U. S. DEPARTMENT OF COMMERCE
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## FORECASTING GUIDE NO. 1

# Forecasting Tornadoes and Severe Thunderstorms

by

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# FOREWORD

During the past several years, the Weather Bureau has given increasing attention to the forecasting of severe local storms. Since the establishment of the Severe Local Storm ("SELS") Forecast Center (now located at Kansas City, Mo.) in 1952, predictions of tornadoes, severe thunderstorms, and concomitant phenomena have been put on a more systematic basis. In its first four years the SELS Center has tested both pre-existing and newly proposed forecasting techniques and has employed the most promising of these in its operating routine. In addition, the staff has been developing new or improved prediction methods. The concentrated efforts of members of this unit and other Weather Bureau forecasters and research people have resulted in substantial advances in the forecasting of severe local storms in the past few years. There is still much to learn, of course, and it is felt that continuing research and experience in the next few years should result in further improvement in severe local storm prediction.

Notwithstanding the continuing evolution of severe storm forecasting methods, it was decided to take time to summarize current techniques for the information of forecasters and other interested meteorologists. The original version of this Guide was prepared by the SELS staff under Mr. D. C. House's direction in February 1955 and given a limited distribution in mimeographed form. With the establishment of the Weather Bureau's Forecasting Guides program in May 1955 it was decided that this treatment of severe storm forecasting could serve as a very suitable basis for a forecasting guide, which would be available to all Weather Bureau forecasters. Mr. R. G. Beebe, who played an important role in the preparation of the original report, was called upon to revise and expand the paper as a Forecasting Guide. Other SELS staff members who contributed to this revised version as well as the original report are: Messrs. F. C. Bates, J. T. Lee, D. S. Foster, J. G. Galway, and B. W. Magor. Acknowledgement is made to Mr. C. F. Van Thullenar for his many suggestions for the improvement of basic forecast procedures and his encouragement in the preparation of the original and revised texts, to Miss Georgina Neubrand for the preparation of diagrams, and to Mrs. L. V. Wolford and Messrs. R. W. Schloemer and J. L. Baldwin of the Office of Climatology for providing up-to-date climatological information for figures 1 - 9 and the accompanying text.

Naturally, some of the methods used by the SELS Center differ from those used by other groups (in the Weather Bureau and elsewhere) concerned with the severe local storm problem, since this is a prediction field where techniques are still mainly subjective. However, an attempt has been made to incorporate or refer to practical ideas or methods of other forecasters wherever appropriate.

**JAY S. WINSTON**
**July 1956**

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# Forecasting Tornadoes and Severe Thunderstorms

## INTRODUCTION

This guide is concerned with the problem of short-range predictions (usually 12 hours or less) of the occurrence of tornadoes and severe thunderstorms over the United States. These constitute what have become known as severe local storm forecasts. Severe thunderstorm forecasts are issued by the Weather Bureau when thunderstorms are expected to result in one or more of the following:

1. Surface gusts of 75 mph (65 knots) or more,
2. Sustained surface winds of 50 mph (44 knots) or more,
3. Hail having a diameter of 3/4 inch or larger,
4. Severe or extreme turbulence.

A tornado forecast implies that the area alerted will also have severe thunderstorms. The treatment in this guide excludes tornadoes and thunderstorms associated with hurricanes, and phenomena such as waterspouts and dust devils.

The size of the severe local storms forecast area is dependent upon the type of weather expected and the season. Areas in which severe thunderstorms are predicted are usually somewhat larger than tornado forecast areas. It has been observed that an area of some 20,000 square miles will usually encompass the severe thunderstorms or tornadoes occurring during a 6-hour period in late winter, spring, and early summer when family-type tornado outbreaks are most frequent. In other seasons (i.e., late summer, early fall) the areas are usually somewhat smaller since tornado occurrences are more isolated in these seasons.

Present severe local storms forecasts are aimed at predicting the areas which are expected to contain the maximum amount of severe weather. Various data restrictions along with the limitations in our current knowledge of the cause of severe local storms, preclude the prediction of a unique area. Thus, each forecast area is surrounded by an area having an expected lower probability of occurrence of severe local storms with the probability decreasing with distance from the predicted area. The same is also true with regard to the valid time of the forecast. It is stressed that present severe local storms forecasts are for areas and no attempt is currently made to predict the time and site of occurrence of any individual severe local storm.

With perhaps a few exceptions, it is generally believed at this time that the difference between severe thunderstorms with tornadoes and those without associated tornadoes is largely a matter of degree. One of the forecasting problems is to evaluate subjectively the various considerations to determine the expected intensity of thunderstorm activity. Thus it is impossible to describe the tornado forecasting method separately from the severe thunderstorm forecasting method without considerable repetition. For simplicity, then, this guide has been prepared in two sections. The first and more comprehensive section deals with tornado forecasting, but naturally includes most general aspects of severe thunderstorm forecasting. The second section treats the specific problems of hail size, turbulence, and surface gusts associated with severe thunderstorms.

## PART I—TORNADO FORECASTING

### HISTORY

In the 1880's Finley [20, 21] studied a large number of tornadoes and was among the first to present detailed information on their development and motion in relation to synoptic pressure patterns on the daily weather map. In addition, he summarized much data on the localized characteristics of tornadoes and on their climatology. Apparently in connection with these studies a tornado forecasting program was in operation, since William A. Eddy, Tornado Reporter of the U. S. Army Signal Service, stated in the October 31, 1885 issue of Scientific American that: "Of 3,228 predictions unfavorable to tornadoes made in 1884, 3,201 were verified..." of 57 predictions that tornadoes would occur in April and June 1884 and June and July 1885, 33 were verified. "When tornadoes were predicted in no instance did violent storms fail to occur, either hurricanes, tornadoes, or hail. The failure of some predictions is doubtless due to inaccurate and insufficient reports from sparsely settled regions." It is presumed that these claims refer to forecasts of a generalized nature since the weather synoptic network stations were widely spaced in the 1880's, the Mid-West population was small, and communications were limited. The facts about what took place in tornado forecasting between the 1880's and 1905 are obscure, but it is believed that little attention was given to the problem of localizing the forecasts primarily because of their complexity. No forecasts are known to have been issued during the late years of the 19th century. In 1905, the Weather Bureau's Station Regulations contained the statement: "forecasts of tornadoes are prohibited." This statement also appeared in the revised Station Regulations issued in 1915 and in 1934. During all of this period forecasters were instructed to predict destructive local storms, but not to mention the word tornado since it was felt that it would cause public alarm and panic. This policy was changed about 1938, but because of the limited knowledge of meteorological processes involved in such forecasting, very few forecasts were issued.

The few kite and airplane observations, beginning in the 1920's, afforded meteorologists the first opportunity to study the air mass in which tornadoes formed. The existence of a temperature inversion in an air mass prior to tornado development was mentioned in an article by Van Everdingen [53] which was summarized by Varney [55]. A little later, Humphreys [28] stated: "Mid-air temperature inversions appear to be quite common and the lapse rates next above these inversions (increase) very rapid (ly), often nearly or quite of adiabatic value. In short, so far as one can infer from these few observations (26 observations by sounding balloon or kite), the atmosphere in the neighborhood of a tornado appears to be unusually stratified, and tending to become unstable at one or more levels." A later article [39], discussing the air mass in connection with a tornado outbreak in Georgia and Alabama, pointed out that the layer of air through the inversion was convectively unstable.

Perhaps the most significant development in the actual prediction of tornadoes occurred in 1942. The April issue of the Monthly Weather Review of that year carried an article, "The Development and Trajectories of Tornadoes", by Lloyd [33] which summarized some study and experience with this problem. This article presented several precedent soundings that were convectively unstable. About the same time, Showalter and Fulks began a study of tornado forecasting problems and their research was summarized in the Weather Bureau's Preliminary Report on Tornadoes [43]. This research emphasized the importance of convective instability, was illustrated with numerous examples of synoptic conditions preceding tornado development, and outlined several suggestions on tornado forecasting. A typical sounding was proposed which was based upon many soundings taken prior to tornado development and within the same air mass. Even though these studies pointed the way towards tornado forecasting (actually providing much of the basis for such predictions until about 1953), few tornado forecasts were issued by the Weather Bureau during the 1940's. The exceptions were those messages sent by the district forecasters at Kansas City, Mo., to the St. Louis Area Headquarters of the Red Cross, alerting their disaster officials to the possibility of tornado developments.

Fawbush and Miller of the Air Weather Service became interested in tornadoes when one resulted in severe damage to Tinker Field in 1948. These two forecasters expanded upon the work of several writers [32, 33, 43, 51] and listed a requirement for the simultaneous occurrence of 6 criteria in areas where tornadoes could be expected to occur. Fourteen tornado forecasts were issued by these two forecasters in 1949 for the Tinker Air Force Base area, and the Severe Weather Warning Center of the Air Weather Service began issuing forecasts of severe weather for the entire United States in 1952.

In March 1952, the Weather Bureau established a Center in Washington, D. C. for the purpose of assisting district forecasters in prediction of severe local storms by preparing special charts as well as by maintaining a continuous watch for areas of possible tornado development. Initially, the forecasts were based largely upon the work of Fawbush, Miller, and Starrett [19] and Showalter and Fulks [43]. The SELS Center (Severe Local Storm Forecast Center) was moved to the Weather Bureau Forecast Center at Kansas City, Mo. in August 1954. This change was made primarily to provide closer contact and more rapid communications among the three forecast centers (Kansas City, Chicago, and New Orleans) whose districts include the major tornado areas.

The functions and responsibilities of the SELS Center remain essentially the same as when established - namely, to provide guidance and assistance to district forecasters in the forecasting of areas in which conditions are favorable for the development of tornadoes and/or severe thunderstorms. During the few years of its operation, however, the SELS Center has developed several new or improved forecasting methods which will be treated in this Guide.

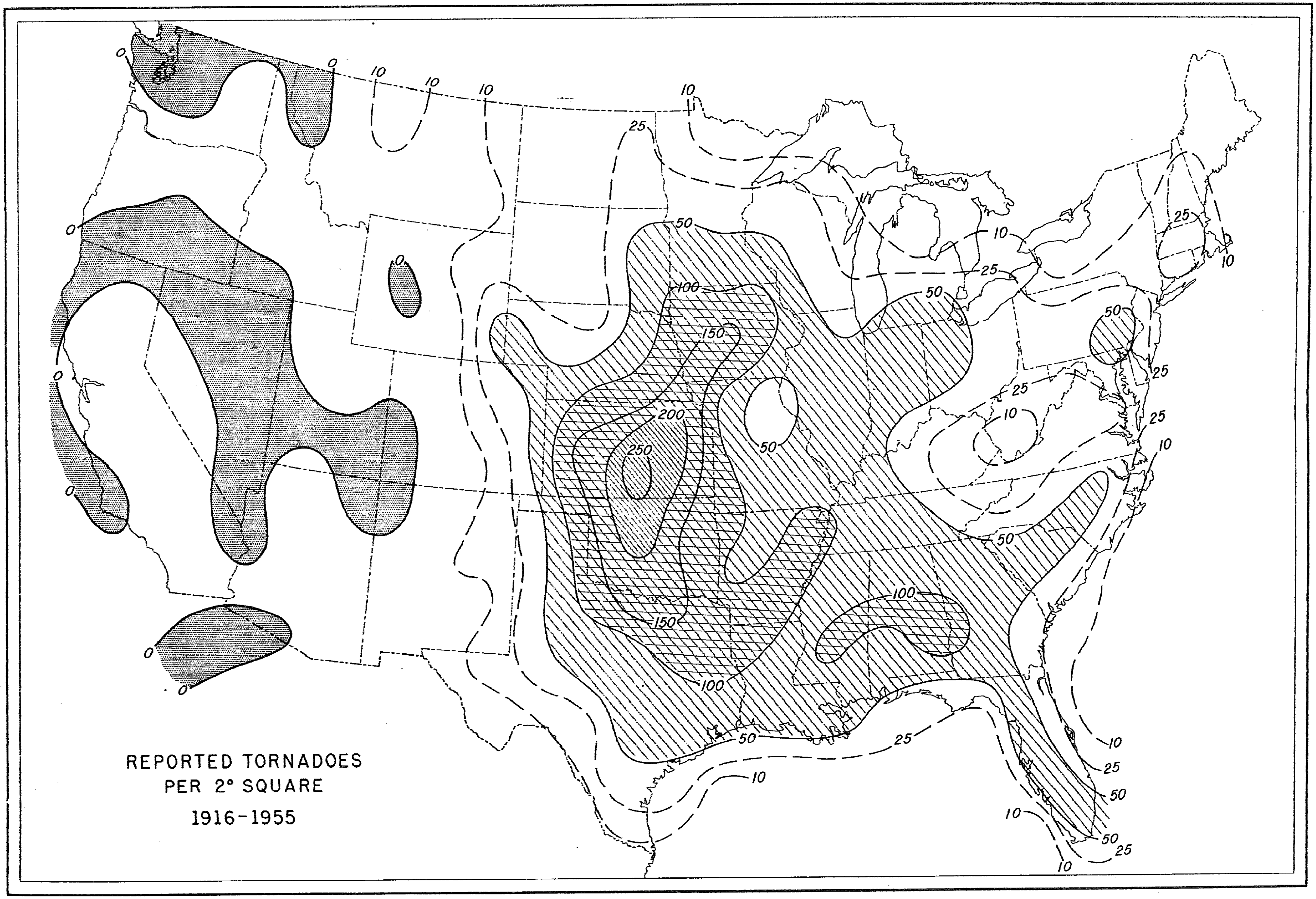

Figure 1. Total number of reported tornadoes per 2° latitude-longitude square over the United States, 1916-1955.

During this developmental stage of tornado forecasting, several objective methods of forecasting tornadoes for specified areas were derived. These include: Armstrong's method for Georgia [2]; Kraft's method for the Gulf States [29]; Mook's method for Ohio [36]; Schmidt's method for western Kentucky and western Tennessee [41]; and the Shuman-Carstensen methods for the Mississippi Valley [45] and for Kansas and Nebraska [44]. While all of these objective methods provided a typing system of the synoptic charts they did not, or perhaps could not, take the thermodynamic features of the various air masses into account except as grossly implied by the synoptic types. Another limitation in the application of these methods is that the size of the areas as well as the time periods involved are entirely too large to be used consistently without adverse public reaction. Tornado occurrence and upper air data available to these researchers were limited so that the normals implied in these studies are not strictly applicable to the much improved reporting system (both with regard to upper air soundings and tornado reports) now in operation. Tests conducted by the SELS Center during the past 2 years have shown that some of these studies (Armstrong and Shuman-Carstensen in particular) held up very well while others did not. Some of these objective methods may be used as a guide regarding the possibility or probability of tornadoes in specific areas, but the indications from these methods must be used cautiously.

## CLIMATOLOGY

### Geographic, Seasonal, and Diurnal Distribution

The importance of the geographic, seasonal, and diurnal distribution of tornado occurrences must not be overlooked in tornado forecasting. Figure 1 shows the occurrences of all reported tornadoes from 1916 through 1955. From this chart, it may be noted that the area of maximum annual frequency is in the Kansas-Iowa area. While tornadoes have been reported in every State, occurrences are rare west of the Continental Divide and practically unknown in the western mountain areas. Charts 1-12 of Technical Paper No. 20 [52] illustrate the distribution by months during the 35-year period, 1916-1950. Of particular interest is the seasonal progression of the area of maximum frequency from the South and Southeast in winter to the Southern Plains in spring and on to the Central and Northern Plains in summer. Figures 2, 3, and 4, obtained from the Office of Climatology, also illustrate this same seasonal progression. Figure 5 shows the monthly distribution of tornado occurrences in the entire United States.

Climatological data on the diurnal distribution by months or seasons are not readily available. Figures 6 through 9, also supplied by the Office of Climatology, show the diurnal distribution of occurrences throughout the year for the entire United States and also for three large regions of relatively frequent tornado occurrence. In figure 6, note that the afternoon maximum for the Southeast from 1300-1900 LST is only about four times that of the early morning minimum from 0300 to 0900 LST. On the other hand, the afternoon maximum in the Oklahoma-Arkansas-Kansas-

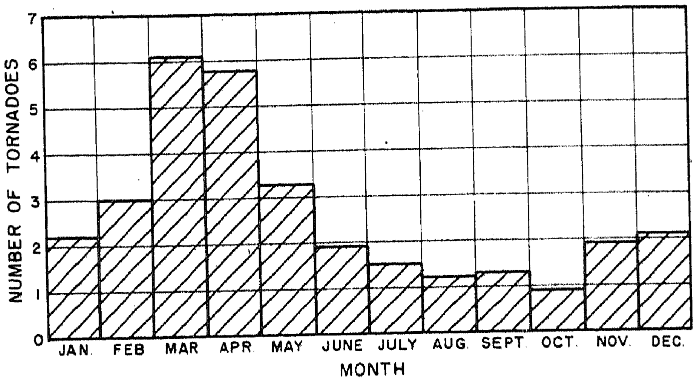

Figure 2. Mean monthly number of tornadoes in the Florida-Georgia-Alabama-Mississippi-Louisiana area, 1916-1955. Based on 1,248 tornadoes.

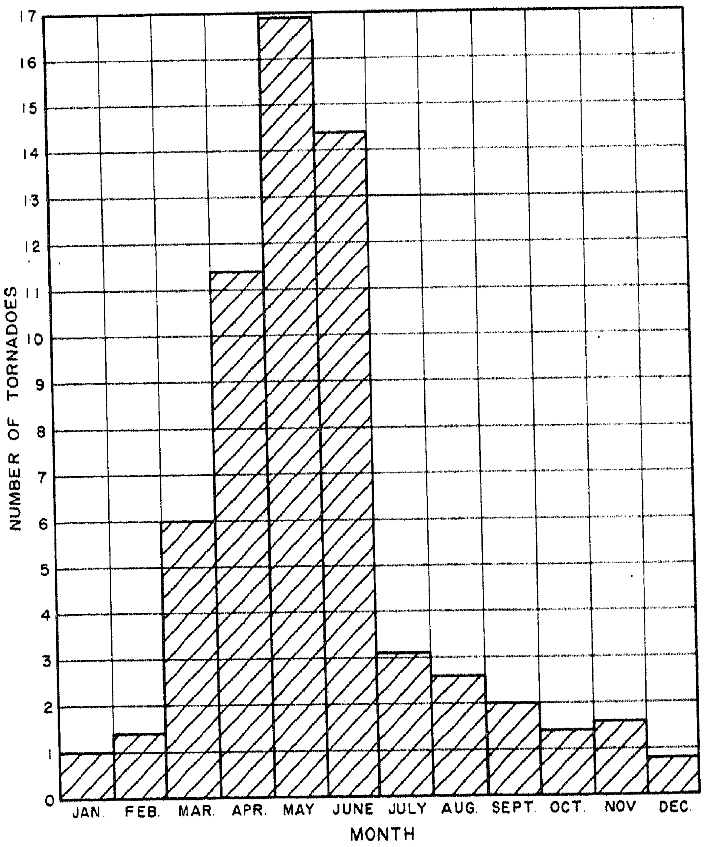

Figure 3. Mean monthly number of tornadoes in the Oklahoma-Arkansas-Kansas-Missouri area, 1916-1955. Based on 2,502 tornadoes.

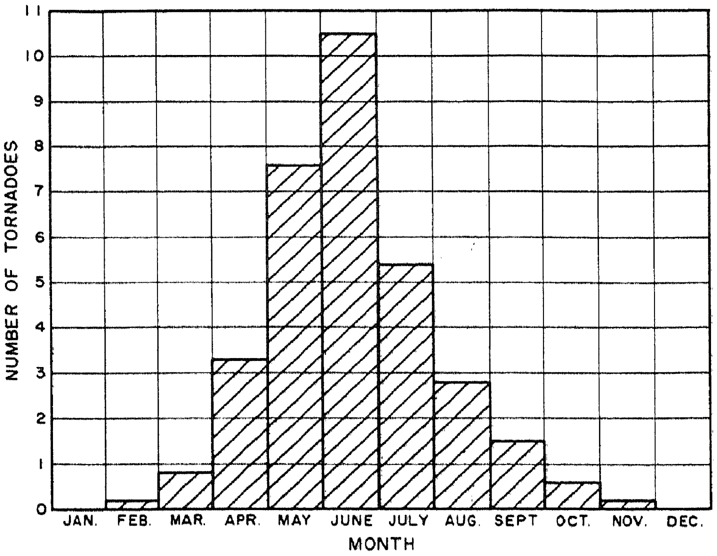

Figure 4. Mean monthly number of tornadoes in the Nebraska-Iowa-Dakota-Minnesota area, 1916-1955. Based on 1,309 tornadoes.

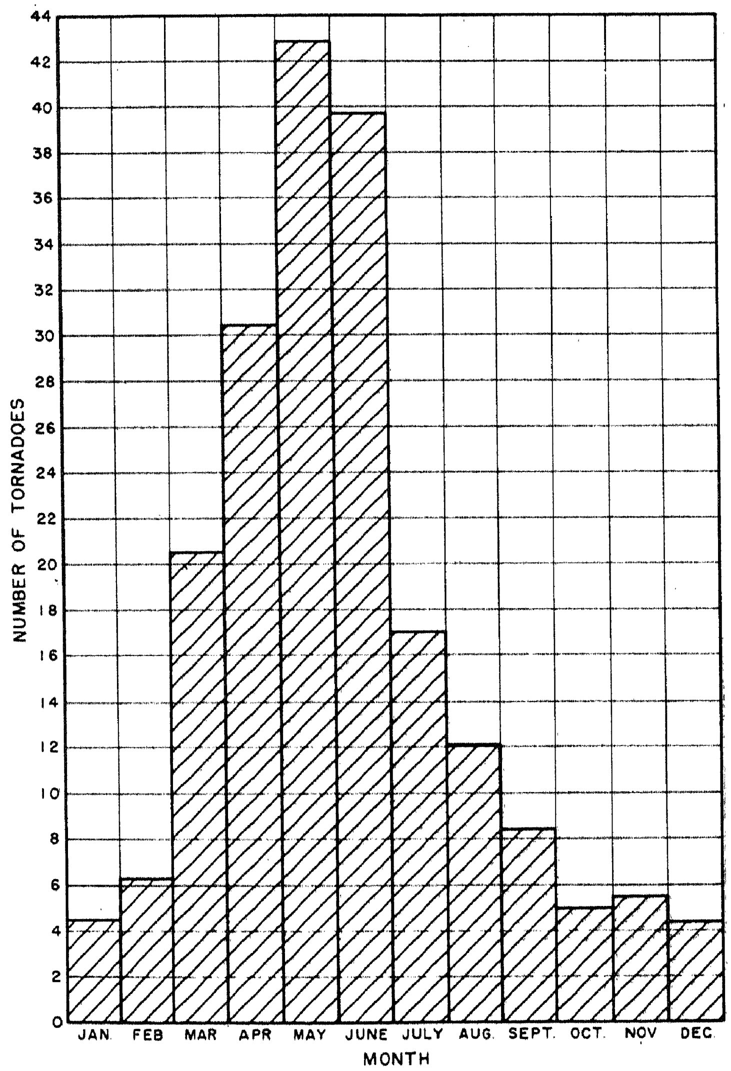

Figure 5. Mean monthly number of tornadoes in the United States, 1916-1955. Based on 7,867 tornadoes.

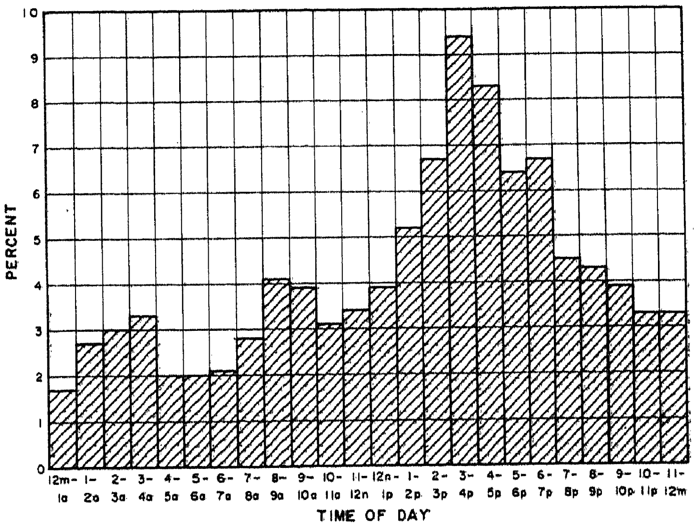

Figure 6. Percentage of tornadoes beginning during the specified hours in the Florida-Alabama-Mississippi-Louisiana area, 1916-1955. Based on 1,127 tornadoes. Time of day not available for the remaining 121 of the 1,248 tornadoes reported in this area.

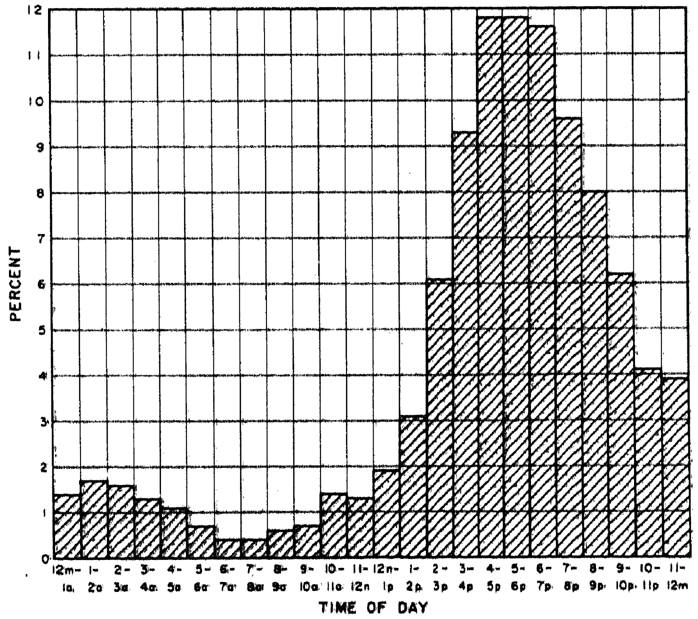

Figure 7. Percentage of tornadoes beginning during the specified hours in the Oklahoma-Arkansas-Kansas-Missouri area, 1916-1955. Based on 2,338 tornadoes. Time of day not available for the remaining 164 of the 2,502 tornadoes reported in this area.

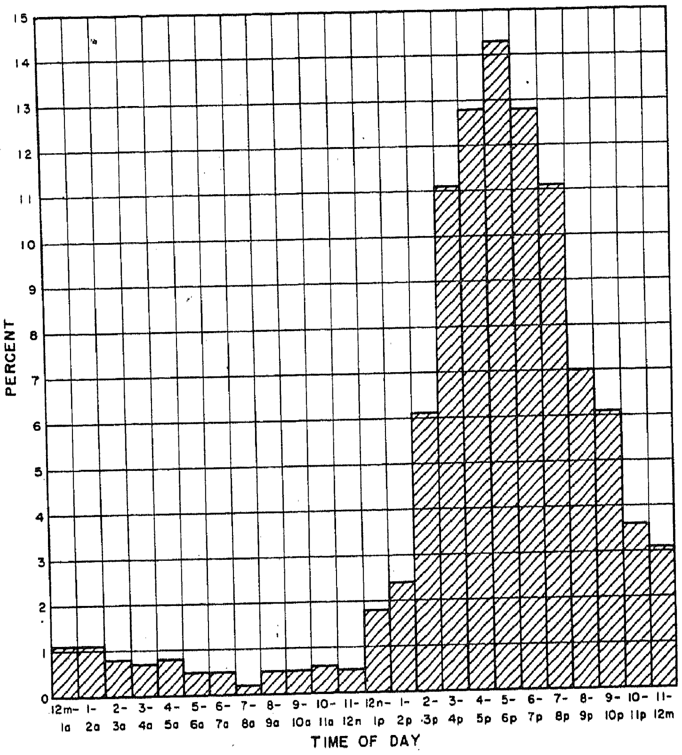

Figure 8. Percentage of tornadoes beginning during the specified hours in the Nebraska-Iowa-Dakota-Minnesota area, 1916-1955. Based on 1,202 tornadoes. Time of day not available for the remaining 107 of the 1,309 tornadoes reported in this area.

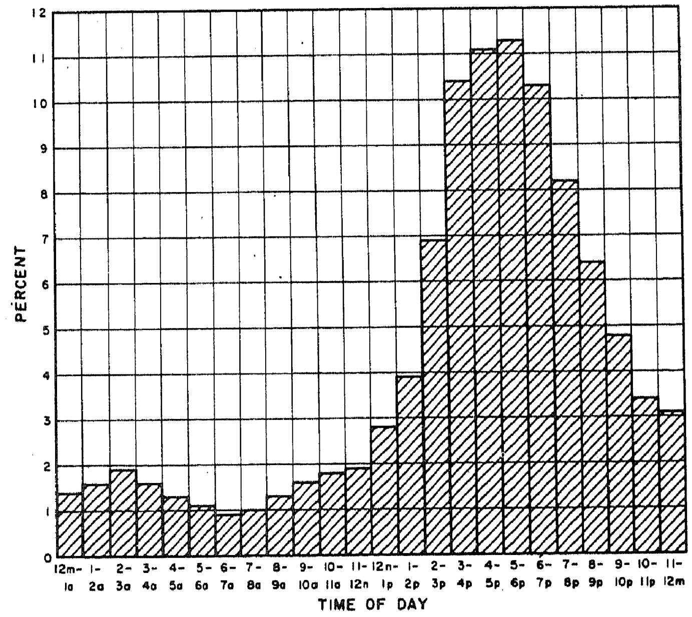

Figure 9. Percentage of tornadoes beginning during the specified hours in the entire United States, 1916-1955. Based on 7,202 tornadoes. Time of day not available for the remaining 665 tornadoes.

Missouri region between 1600-1900 LST, shown in figure 7, is about 25 times greater than during the minimum period from 0600 to 1000 LST. This difference is even more pronounced for the Nebraska-Iowa-Dakota-Minnesota area as shown in figure 8. Thus it is evident that the diurnal distribution of occurrences is much more important over the Central and Northern Plains than over the Southeast.

The climatological expectancy of a tornado occurring within a specific 20,000-square-mile area, within a given month, and during a particular 6-hour period never even approaches the 50% probability level. For example, consider the climatological probability for Kansas (82,000 sq. mi.) during the month of maximum frequency, June (8.1 tornadoes per month), and during the period of maximum diurnal frequency, 1500-2100 CST (68% of all tornadoes in Kansas occur during this period). The climatological expectancy is as follows:

20,000/82,000 x 8.1/30 x .68 = .04 ,

or a 4% probability of verification if the day and area are selected at random within the limits specified above. During the past several years more tornado reports have been logged as compared with the remainder of the 40-year period from 1916 through 1955. Even so, the highest climatological expectancy of a tornado forecast area being verified (Kansas) is probably less than 10%. The expectancy in Alabama in March is about $\frac{1}{2}$ of 1%, based upon data from 1916 through 1955. Considering the area east of the Continental Divide, all months, and four 6-hour periods per day, the climatological expectancy of a tornado occurrence within a 20,000-square-mile area during a 6-hour period is about 1/400. Thus, the tornado forecaster must always deal with situations in which the climatological chance of verification is much lower than is true of most types of weather forecasting.

During the first 7 months of 1955, verification records showed that 35%, or 1 in 3, of all SELS tornado forecast areas (averaging 20,000 sq. mi. in 6 hours) contained one or more tornadoes (based upon unofficial tornado reports) during the valid time. Increasing the areas by 150 miles in each direction and adding 2 hours to the valid time, 59% (including the 35% listed above) of these enlarged forecast areas contained 1 or more tornadoes. While the expectancy of verification of a tornado forecast, 1 in 3, seems unimpressively low, this has been attained against a climatological expectancy of 1 in 400. By comparison, 12-hour precipitation forecasts in most sections of the United States east of the Continental Divide may be expected to verify about 1 time in 2 occurrences. The climatological probability of verification of precipitation forecasts during 12-hour periods varies from about 1 in 4 to 1 in 6.

## Air Mass Types

Much research on tornado forecasting has dealt with a "typical" [43] or "mean" [15] sounding as representative of the air mass in which tornadoes occurred. Such soundings are characterized by a dry layer overlying a moist layer near the surface as well as both conditional and convective instability. More emphasis on tornado forecasting during the past few years has resulted in the procurement of many more upper air soundings during the tornado season than were available for study heretofore. Many soundings obtained just prior to or associated with tornado occurrences have been observed to differ markedly from the "typical" or "mean" soundings. A preliminary study in the SELS Center [5] of most of the tornado occurrences in the United States during 1952 and 1953, plus other interesting cases, pointed to 4 broad types of soundings as representative of the associated air mass in which tornadoes form. A study yielding similar results was conducted by the Air Force Severe Weather Warning Center [18].

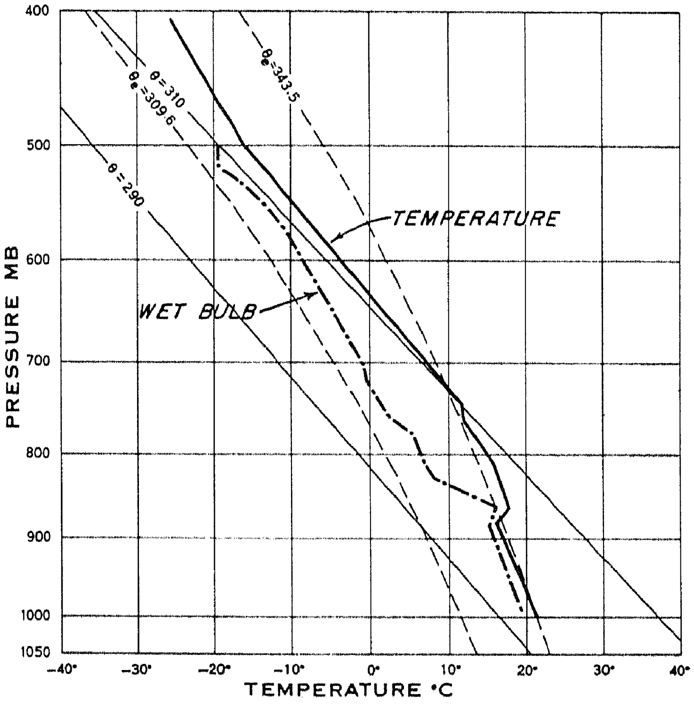

Figure 10. Example of Type I upper air sounding, Barksdale AFB, La., 0900 CST, March 21, 1952. The severe outbreak of tornadoes over Arkansas and Tennessee on March 21-22, 1952 began 6 hours after this sounding and the first tornado in this series was reported 100 miles north of Shreveport, La. Tornadoes with this type of precedent sounding most often occur in the South during late winter and early spring, and over the Midwest during spring and early summer.

Examples of these types of upper air soundings preceding tornadoes are given in figures 10 through 13. Precedent soundings have been defined [5] as those characteristic of the air mass but removed in time and/or space from the vicinity of tornado occurrence. It is emphasized that this identification of types is somewhat tentative. Also while a Type IV sounding has been identified, it is not certain at this time whether the tornado actually occurs within this air mass or whether it occurs in conjunction with this air mass and that shown as Type I.

## Tornado Proximity Soundings

The above study also showed, in the case of the Type I air mass, that as the time and site of tornado occurrence was approached, the "typical" inversion gradually disappeared and the moist layer became increasingly deep. Thus it is necessary, when attempting to define or study tornado soundings, to first set some limit in time and/or space since the environment in which the tornado forms is undergoing what appears to be a continual change. On the basis of the foregoing, a tornado proximity sounding was then defined as one in which all of the following criteria are met:

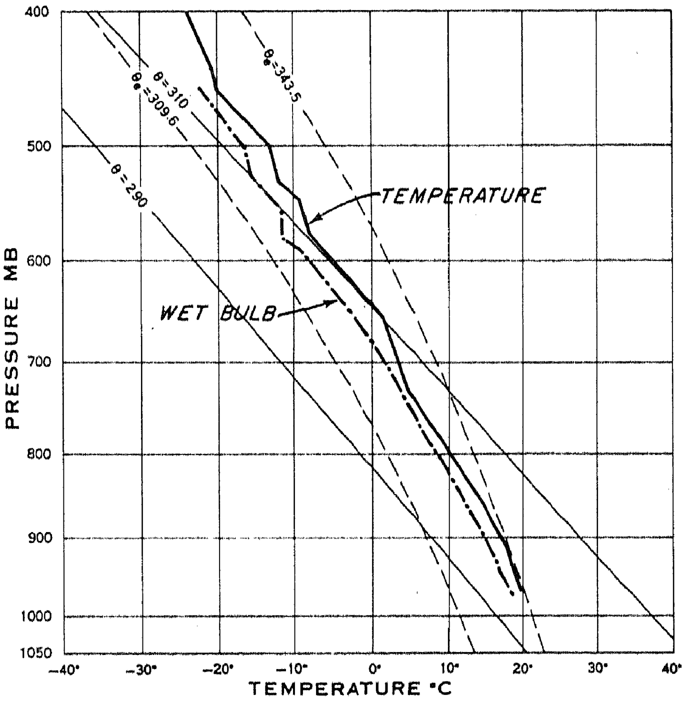

Figure 11. Example of a Type II upper air sounding, Atlanta, Ga., 2200 EST, March 27, 1948. Tornadoes occurred about 60 miles northwest of Atlanta 4 hours after this sounding. This warm, moist, precedent sounding is most often noted in connection with tornadoes in the Southeast and also with tornadoes associated with hurricanes. Occurrences in this type of air mass are rather infrequent.

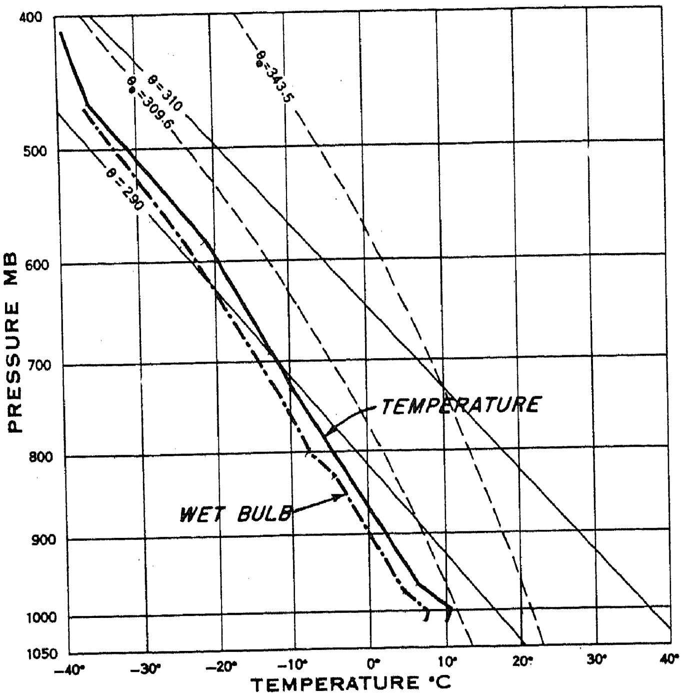

Figure 12. Example of a Type III upper air sounding, Mitchel AFB, N. Y., 1000 EST, May 10, 1954. Tornadoes occurred 100 miles north of Mitchel AFB about 3 hours after this sounding. Occurrences in this cold, moist air mass are also rather infrequent and often consist of reports of funnel clouds aloft rather than destructive funnels reaching the ground.

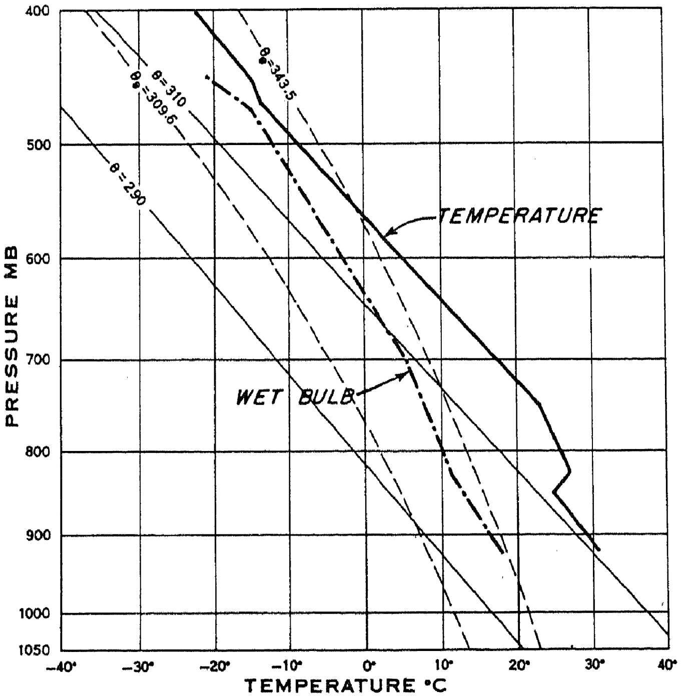

Figure 13. Example of a Type IV upper air sounding, Dodge City, Kans., 0900 CST, June 24, 1953. A tornado was reported 45 miles south-southeast of Dodge City 7 hours after this sounding.

1. Tornado occurrence within 50 miles of a raob station.
2. Tornado occurrence within the hour following release of the radiosonde instrument.
3. Tornado occurrence within the same air mass as that in which the sounding was taken.
4. Sounding taken ahead of or near the parent thunderstorm but not behind it.

A proximity sounding thus represents conditions over a relatively small area near the tornado, and comparison with precedent soundings reveals the changes in the environment associated with severe storm development. A proximity sounding then is not necessarily characteristic of the air mass itself. A proximity sounding of the Type IV air mass is shown in figure 14.

In a study of tornado proximity soundings [6], 11 examples were found wherein the precedent air mass was classified as Type I. An inspection of

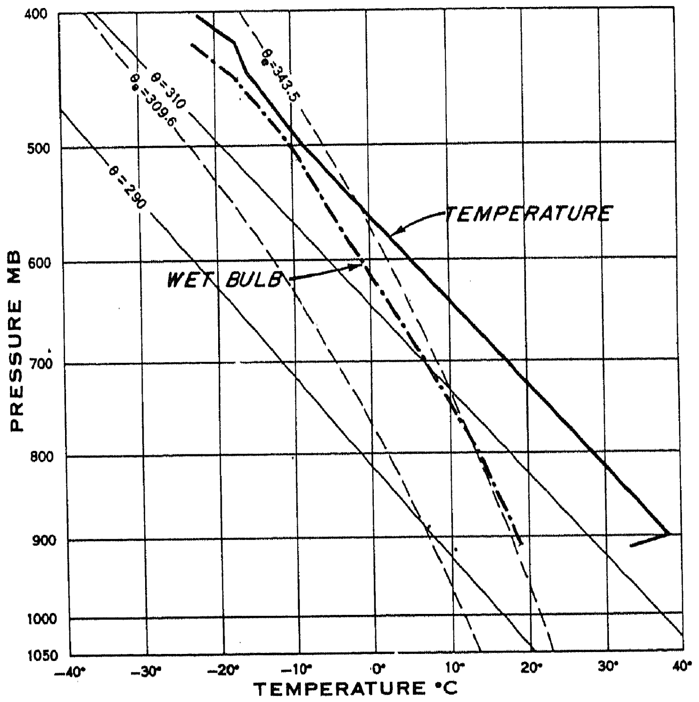

Figure 14. Example of Type IV upper air sounding (6 hours later than fig. 13), Dodge City, Kans., 1500 CST, June 24, 1953. A tornado was reported 45 miles south-southwest of Dodge City one hour after this proximity sounding. This type of sounding (termed the "inverted V" in the SELS Center) is noted in connection with many of the long, narrow, rope-like funnel clouds that occur during the summer months over the higher Plains States. Of particular interest here is the complete absence of either a stable layer or a pronounced lower moist layer. The diurnal variation in these occurrences is very pronounced with the maximum occurring near, or just after, the normal time of maximum temperature.

these 11 soundings revealed that none of them contained a "typical" inversion, but all were characterized rather by the penetration of the moisture to great heights (averaging 16,000 feet above the surface). All of these 11 soundings were both conditionally and convectively unstable. These soundings were analyzed to determine the various parameters commonly used in severe local storm forecasting. One of the most interesting features brought out in this analysis was the change in depth of the moist layer over the area in which tornadoes developed. The moist layer has no precise definition, but the top of the moist layer is usually considered as that point where the moisture (either dew point or wet bulb temperature) shows a marked decrease with height.

Using data from these 11 cases, a composite chart was prepared by plotting moist layer depths relative to the point of the tornado occurrence. Data from 120 soundings were used in the preparation of this composite chart, shown in figure 15. It appears that tornadoes actually occur within areas of deep moist layers which are surrounded by lesser depths.

At the same time that the moist layer deepens over the areas where tornadoes develop, it becomes more shallow to the south in many cases. Figure 16 is an example of a proximity sounding (a tornado occurred 20 miles south-southeast of Oklahoma City about 1 hour later), while figure 17 is an example of a Type I precedent sounding (6 hours before). In this case, the depth of the moist layer increased from 2,700 feet to above 13,000 feet (humidity data above this level were missing). At the next raob station to the south,

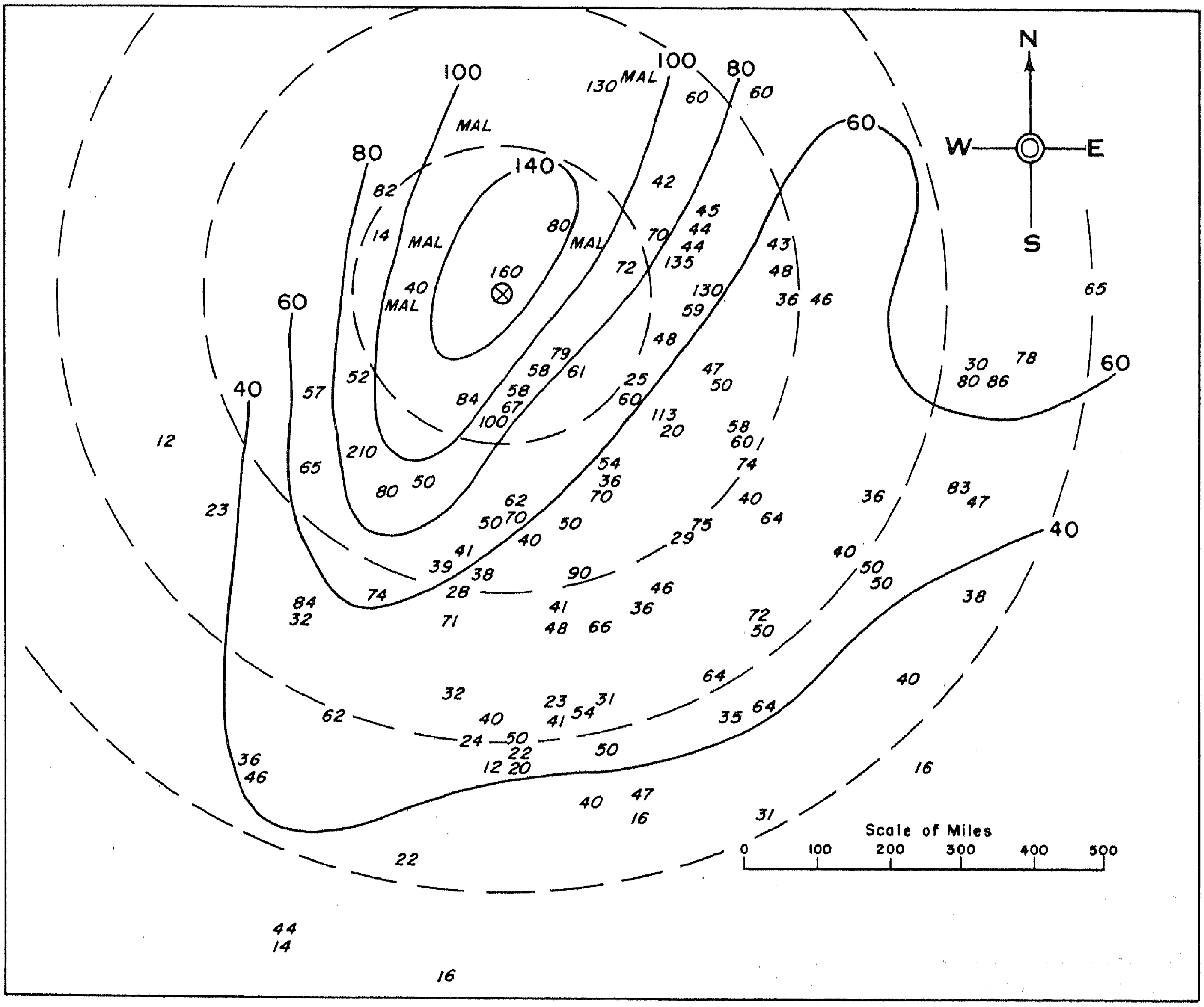

Figure 15. Composite chart showing depths of the moist layer in eleven Type I cases. Isopleths of the average depth of the moist layer in hundreds of feet for these cases are shown. ⊗ indicates point of tornado occurrence, and dashed radial lines are drawn at 200-mile intervals.

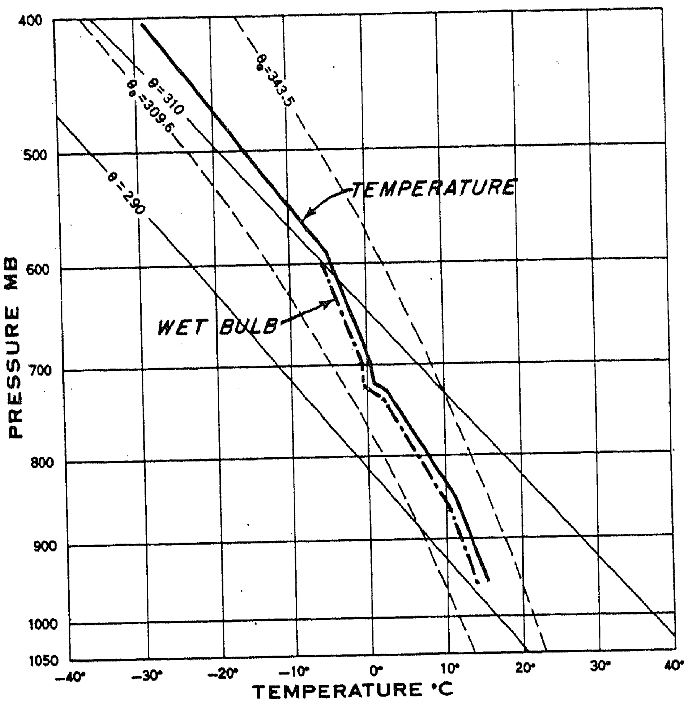

Figure 16. Upper air sounding at Oklahoma City, February 19, 1951, 2100 CST. A tornado was reported 20 miles south-southwest of Oklahoma City one hour later.

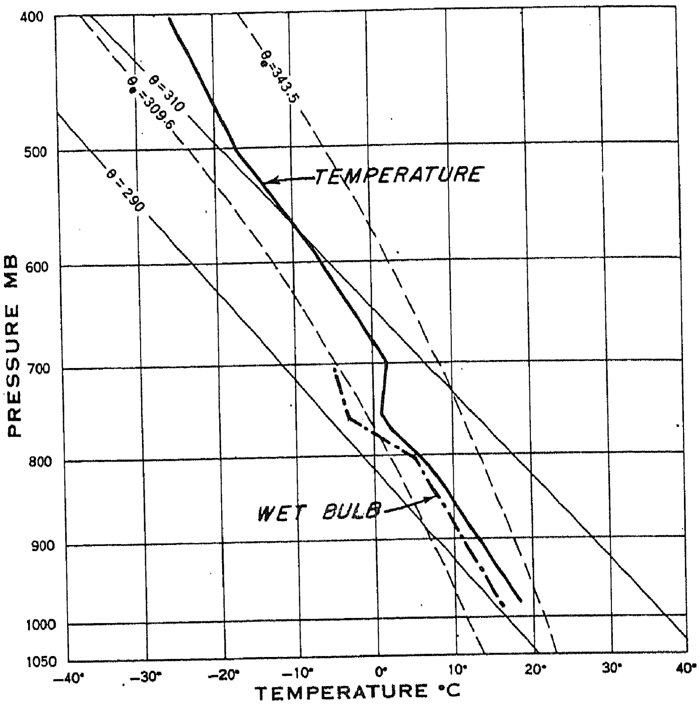

Figure 18. Upper air sounding at Fort Worth, Tex., February 19, 1951, 2100 CST.

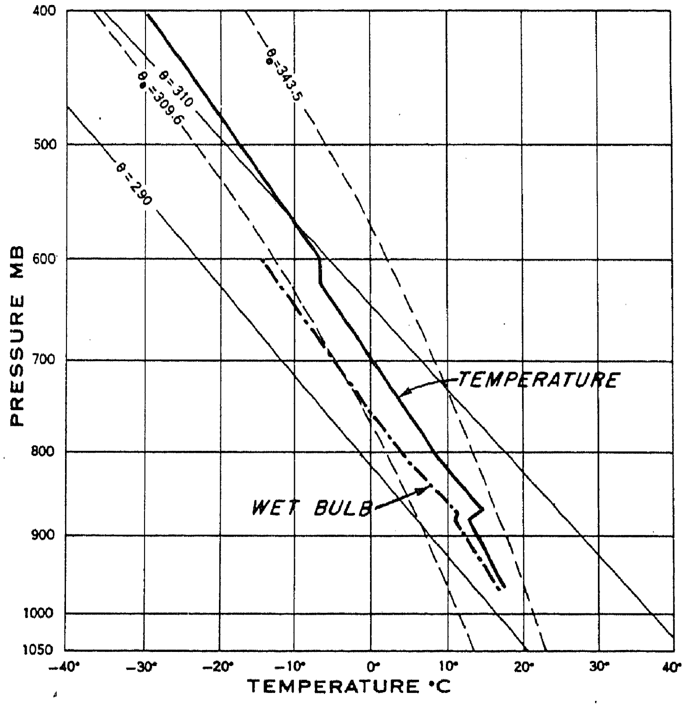

Figure 17. Upper air sounding at Tinker AFB, Okla., February 19, 1951, 1600 CST.

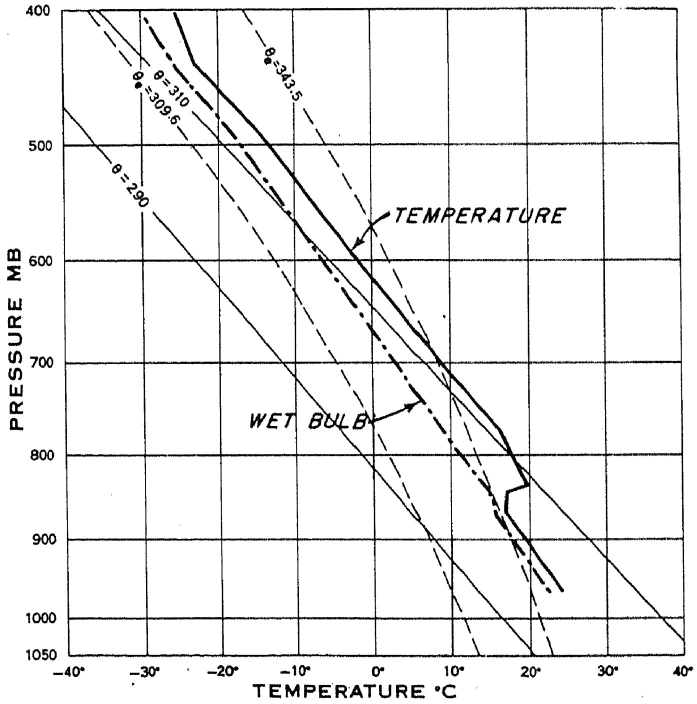

Figure 19. Upper air sounding at San Antonio, Tex., February 19, 1951, 2100 CST.

Fort Worth, the depth of the moist layer at 2100 CST (fig. 18) was nearly the same as it had been 12 hours earlier except that the "typical" inversion had become even more pronounced. Meanwhile, the depth at San Antonio (fig. 19) had decreased with marked drying aloft as compared with 12 hours earlier (not shown). Farther to the south, at Brownsville, the depth was only 2,700 feet at

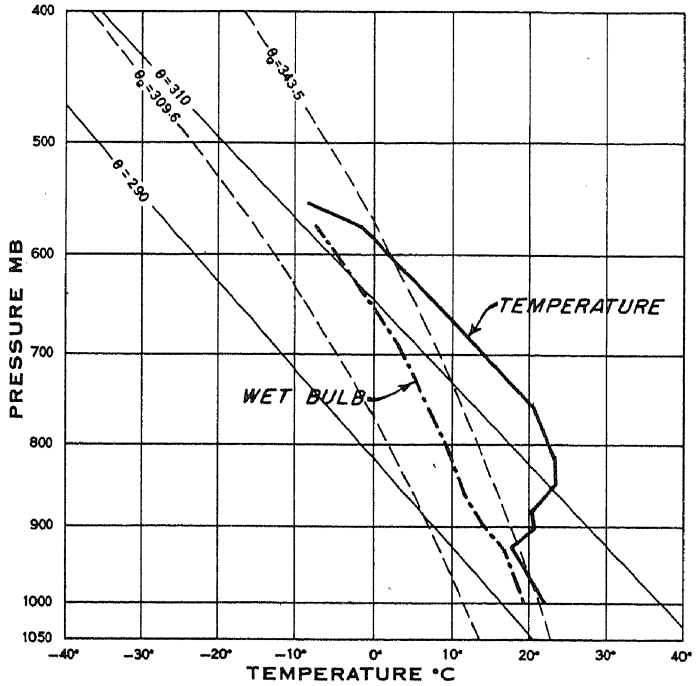

Figure 20. Upper air sounding at Brownsville, Tex., February 19, 1951, 2100 CST.

2100 CST (fig. 20) having lowered from 4,700 feet 12 hours earlier. Thus, the moist layer which initially was of a more or less uniform depth increased markedly with time in the vicinity of tornado formation, while simultaneously it lowered markedly to the south.

From the study and analysis of these data, the following conclusions have been drawn:

1. Data available at this time clearly show that at the time and site of tornado occurrence, the "typical" inversion and presence of relatively dry air aloft are not characteristic of the air mass types investigated here.
2. Tornado forecasting in connection with the Type I sounding must take into account the manner in which the conversion from the precedent sounding to the proximity sounding is effected.
3. Attempts to utilize forecast parameters based upon the presence of the "typical" inversion and/or relatively dry air aloft at the time and site of tornado occurrence may be misleading.

# FORECASTING METHODS

## Synoptic Considerations

It is currently assumed that the tornado develops through the conversion of potential to kinetic energy accompanying the release of instability. Tornadoes are nearly always associated with thunderstorms so that the tornado forecasting methods to be described here are directed first towards the prediction of thunderstorm areas. The area in which tornado occurrence is most likely is that where the most severe thunderstorms are predicted, with the probability of tornado occurrence proportional to the severity of thunderstorms. There are, of course, a few cases of tornadoes which are not associated with thunderstorms and satisfactory methods for predicting these occurrences have yet to be developed. It follows then that tornado forecasts are not currently issued for areas in which thunderstorms are not expected. However, this does not mean that a forecast for tornado occurrence, other conditions being favorable, should be held up until the development of the first shower or thunderstorm. There are numerous cases in which a tornado, or even the first of a series, occurred with the first thunderstorm.

The surface map is generally given first consideration in evaluating the tornado threat of a particular situation. It is agreed that tornadoes most often occur in connection with low centers which are already quite intense or are currently deepening.

Most tornadoes are found within the southeast section of a low and within, or near, the warm sector [9, 10, 23, 24, 29, 30, 43]. These occurrences may be at intersections of warm fronts and instability lines, along instability lines, along cold fronts, and occasionally near a deep low center with no pronounced front evident. Development or intensification of low centers is therefore of extreme importance in this type of forecasting. Cyclonic development may vary from the rapid and intense deepening observed in winter and early spring when spectacular family-type outbreaks occur, to smaller-scale intensification of weak low centers over the Plains States in summer. Although intensity of development varies considerably, it should be pointed out that the deepening referred to here is not the same as the development of the tornado low itself as described by Brooks [9], Means [35], and more recently by Tepper and Eggert [50]. Magor [34] has found several situations wherein a low, whose size is somewhere between that of the tornado low and the average synoptic low center, and which is distinct from the primary low, can be followed for several hours with the present network of synoptic reporting stations.

Researchers generally agree that surface dew points are high at the time and site of most tornado occurrences. Brooks [10] has pointed out that most (111 out of 142) tornadoes "occurred on the

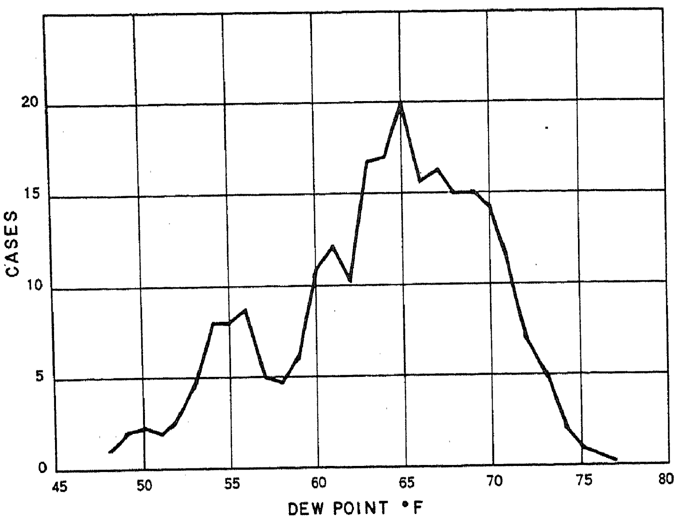

Figure 21. Frequency distribution of 258 tornado occurrences during 1953 in relation to surface dew point.

line of highest surface dew point." Kraft and Conner [30] have suggested occurrences "within 300 miles of 60°F dew point (usually near)." Asp [3] found, in a study of 58 tornadoes in Oklahoma from 1950 through 1952, that tornadoes occurred with surface dew points as low as 31°F and as high as 76°F but 88% of the occurrences were with dew points of 55°F or greater. Fawbush, Miller and Starrett [19] have stated: "Provided that the other criteria are satisfied, tornado development will occur first on the windward border of the moisture wedge near the 55°F dew point isotherm." Beebe [7] in a study of composite tornado charts based upon tornado outbreaks found the surface dew point in the centroid of the area to be near 60°F, with only small differences from one geographic location to another. Galway [24] has made the most comprehensive study of this relationship and his frequency distribution of tornado occurrences according to dew point values is shown in figure 21. It may be noted that over 90% of all occurrences are with dew points having a value of 53°F or greater. This is the critical value in general use now.

At the 850-mb level, cyclogenetical situations are considered to be of primary importance. One of the favorable indications for cyclonic development, as pointed out by George [26], is a strong injection of cold air behind the trough line. This is usually associated with strong southerly winds ahead of the trough, warm air advection and an accompanying or developing temperature ridge, and a strong influx of moisture. The term moisture injection is currently used to depict a southerly wind jet at this level within an area of high moisture values. While the low-level moisture usually extends through the 850-mb surface, a lower limit on the depth of the moist layer has been tentatively set at 3,000 feet. Thus, there are some cases in which the soundings must be inspected or analyzed to ascertain moisture values since low values of the dew point at the 850-mb level do not necessarily indicate "dry air." In any case, it is imperative that a strong influx of moisture be in evidence in tornado situations. Otherwise, when cumulonimbus clouds develop and the moisture is distributed through a depth comparable to the vertical extent of the clouds, the low-level supply would be appreciably lowered, the level of free convection would increase to greater heights, and convective activity would be of only transitory duration. Therefore, the location and position of the low-level moisture injection is one of the more important factors in severe local storms forecasting.

Some tests have been made by the SELS Center wherein areas of tornado occurrence were compared with those areas of maximum horizontal temperature advection (relative warming) at the 850-mb level as developed by Gilman [27,1]. This method, while not originally intended as an aid in tornado forecasting, does provide a measure of low-level horizontal convergence and it has been shown to be of considerable help in the forecasting of heavy rain. Tests have shown that this method usually delineates a small area which is nearly coincident with the tornado areas when the time of tornado occurrence is known. While these tests were most encouraging, the procedure has three drawbacks, which have precluded the adoption of it on a routine basis as far as present SELS operations are concerned. These are: 1. the lack of upper air data over the Gulf of Mexico; 2. the difficulty of standardizing at a single level since there are large differences in the elevations at which tornadoes occur over the United States; and 3. the timing problem. Study on the resolution of these problems is continuing.

The other upper level charts should also indicate deepening if tornadoes are to be predicted. Of particular importance at the 700-mb level is the presence of a trough to the west of a threat area with cold air advection behind the trough and advancing through the trough line. Likewise at the 500-mb level, a trough is generally expected west of a threat area and cold air advection should be predicted at least near, if not over, the tornado threat area.

The following guidance list is intended to summarize synoptic conditions which are generally considered to be favorable for tornado development or to delineate areas in which tornadoes occur. However, it is emphasized that, in particular instances, all of these conditions are neither necessary nor sufficient. Reasons for some of these associations will be dealt with later.

1. Surface Map
   a. The warm sector of a wave cyclone.
   b. Intersections of warm fronts and instability lines.
   c. Instability lines.
   d. Cold fronts.
   e. Areas having a surface dew point of 53°F or greater.
2. 850-mb Chart
   a. Trough to the west of the area.
   b. Moisture injection.
   c. Warm air advection.
   d. Coincidence, or near-coincidence of temperature and dew point ridge.
3. 700-mb Chart
   a. Trough to the west of the area.
   b. "No-change" lines. (See description.)
   c. Areas with increasing local change of moisture values are most favorable.
4. 500-mb Chart
   a. Jet intersections with low-level jet or low-level convergence region. (See description.)
   b. Evidence of dynamic cooling (i.e., cooling not accounted for by horizontal advection).
5. Radar Reports

   At the present time little has been confirmed as to the appearance of a tornado or severe local storm on the radar scope. The following echo patterns are suspected of being indicative of severe local storm activity:
   a. Echoes which indicate vortical shear (hooks or 6-shaped echoes, s-bands, crescent-shaped echoes, small holes in strong cells, etc.)
   b. Intersection of two or more lines of echoes.
   c. Areas of maximum vertical development (cells with tops 40,000 feet or more).
   d. Areas of marked convergence of echo paths.
   e. Well-defined lines of strong convective echoes or very bright echoes just ahead of a well-defined line of strong echoes.
   f. Unusually strong isolated cells, especially those moving at a high rate of speed.

During the 1955 tornado season radar reports were frequently of considerable value in extending the time and/or location of tornado forecasts after the convective activity was under way. Also, there are situations wherein the forecaster is undecided as to exactly where an instability line will develop and/or whether convective instability will be released. In such situations radar reports may be invaluable in indicating where and/or when activity is developing. Normally, the beginning of thunderstorms should be forecast through other means since the time interval between raindrop formation, as seen by radar, and severe thunderstorm or tornado occurrence may be very short. The optimum use of radar remains on the local station level where the proper interpretation of the scope patterns can be of inestimable value in local alerting and warning procedures.

## Thermodynamic Considerations

Stability Discussion.—While discussions of stability and its use in forecasting may be found in many meteorological textbooks, there are sometimes slight differences in terms and definitions. So the following discussion is a brief review of stability with emphasis on the use of plotted soundings on pseudoadiabatic charts:

1. Absolute stability (frequently termed simply "stability") is characterized by a temperature lapse rate that is less than the moist adiabatic rate. A parcel moved upward or downward through this layer would, because of buoyancy forces, tend to return to its original pressure, regardless of whether the parcel is saturated or not.

2. Absolute instability is characterized by a temperature lapse rate that is greater than the dry adiabatic rate. A parcel moved upward in such a layer would continue to rise of its own accord and accelerate since the parcel temperature would be warmer than its environment. Similarly, a parcel displaced downward in a layer that is absolutely unstable would continue to sink and accelerate since the parcel temperature would be colder than its environment.

3. Neutral equilibrium is characterized by a lapse rate that is coincident with the dry adiabatic rate (for non-saturated air) and the moist adiabatic rate (for saturated air). A displaced parcel would have the same density as its environment throughout the layer, providing the parcel remains non-saturated or saturated as the case may be. Thus the atmosphere would neither assist nor retard vertical motions in this case.

4. Conditional instability is characterized by a temperature lapse rate which lies between the dry and moist adiabats (i.e., a lapse rate which is steeper than the moist adiabatic lapse rate, but not as steep as the dry adiabatic lapse rate). It should be noted that the moisture content of the layer is not a factor in the definition. However, saturation or non-saturation determines whether the lapse rate is unstable or stable, respectively. Research on tornado forecasting has shown that the lapse rate from near the surface to 500-mb is generally conditionally unstable near the time and site of tornado occurrence.

5. Latent instability, as defined by Petterssen [38], refers to parcel instability that is latent in

the stratification and which may be released through lifting. This is usually evaluated qualitatively by the parcel method from a sounding plotted on a pseudoadiabatic diagram. Latent instability is present whenever a positive area (area above the level of free convection, between the environment lapse rate and the moist adiabatic along which a lifted parcel would travel) is revealed by such an analysis. However, it must be realized that the energy represented by the negative area (area below the level of free convection, between the environment and assumed parcel temperature) must be supplied to effect the lifting before the energy represented by the positive area can be released. Thus the positive area yields energy only after the negative area has been completely overcome.

6. Convective instability may be defined as a condition in which the moisture distribution through a layer is such that the equivalent potential temperature or wet bulb potential temperature decreases with height. It is instability which will be released if the layer is lifted far enough (at least until the base of the layer is lifted beyond its condensation level). This is so because the lapse rate steepens rapidly when the top of the layer cools at the dry rate and the bottom at the moist rate. Convective instability often exists in layers where the temperature lapse rate displays marked stability (inversions), a condition which ordinarily hinders convection.

A convenient means of determining convective instability is from a plot of wet bulb temperatures on a pseudoadiabatic diagram. Convective instability exists if the wet bulb temperature decreases with height more rapidly than the moist adiabatic rate. The importance of convective instability in tornado forecasting is that rapid steepening of the lapse rate may result from lifting by fronts, orographic barriers, low-level horizontal covergence, or pressure jumps.

Showalter Stability Index.—This index [42], which is transmitted regularly both via teletypewriter and facsimile, has proven to be a very useful tool in severe local storm forecasting. Although it is a very simple parameter, it does contain a considerable amount of information about the air mass structure. The Showalter stability index is defined as the observed 500-mb temperature minus the temperature of an air parcel lifted dry and moist-adiabatically from 850 mb. It is thus a first approximation to the estimation of latent instability. Negative values of this index indicate that a positive area exists (for the parcel lifted from 850 mb) and that the layer between 850 and 500 mb is convectively unstable. Positive values indicate lalent stability at least in a gross sense, but do not give any definite information about convective instability. However, for the range of 500-mb temperatures ordinarily encountered it is usually safe to assume that the layer (850 to 500 mb) is convectively stable when the stability index is greater than about +6°. For smaller positive or zero values the index does not uniquely define convective stability or instability.

Thus, negative or small positive values of this index may be considered as a normal necessary condition prior to the development of severe local storms (there are occasional exceptions), but it should be noted, as Showalter has stated, that such values certainly do not constitute a sufficient condition. Plotted values of this index on a map will very quickly and easily delineate most suspect areas and this is the principal use currently made of the stability index. Also, this index, when used in conjunction with the level of free convection, is a tool for making a rapid check upon thunderstorm potential but not for making the forecast itself. There are some severe local storm outbreaks which may not be indicated by a typical analysis of static, low-valued areas of the stability index. The forecasting problem is one of anticipating the time rate of change of stability.

Experience has shown that instability areas, as indicated by this index, cannot usually be extrapolated. Perhaps the major reason for this lies in the arbitrary use of the temperature and dew point at 850 mb as representative of lower-level moisture. Moisture in tropical air masses over the United States often cuts off rapidly with height (e.g., the Type I precedent sounding shown in fig. 10). Thus, when the top of the moist layer moves through the 850-mb surface, radical changes in the index occur. Some study of this problem has resulted in the tentative adoption, in the SELS Center, of the "lifted index" (to be described later) which overcomes this difficulty. A statistical check by Galway [25] has shown that the lifted index is an excellent predictor of the Showalter stability index.

Analysis of Soundings.—The primary objective of the analysis of soundings is the determination of the degree of hydrostatic instability that will be available at the time and site of severe thunderstorm or tornado occurrence. This analysis is carried out by the parcel method, in which the following assumptions are made:

1. An undisturbed environment - no compensating vertical motions resulting from parcel displacement.
2. No horizontal convergence or divergence in the region under consideration.
3. Temperature changes are adiabatic (both dry and moist).
4. No entrainment.

It is obvious that these assumptions are not realistic. The environment cannot remain unchanged as an ascending current must be compensated for by a descending current. Also, the area in which the development of strong convective clouds would be expected is one characterized by low-level horizontal convergence. Entrainment generally occurs in most convective developments, but it is believed that its influence at the core of strong updrafts of large cross-section (the kind that are associated with severe local storms) must be small. It should be pointed out that convergence and entrainment have opposite effects on the accuracy of the parcel method. In the case of low-level horizontal convergence the parcel method underestimates the instability, while in the case of entrainment it overestimates the instability. Since the method is so simple and refinements to it can only be estimated qualitatively, it is probably best to apply it without modification, at least in the first approximation.

In practice the sounding is analyzed by assuming a representative parcel, i.e., obtaining the average temperature and mixing ratio for the layer of air adjacent to the surface. Experience thus far indicates that the lower 3,000-foot layer is most useful for these evaluations. It is also assumed in this sounding analysis that the positive area need not be evaluated any higher than the 400-mb level.

The steps followed in the SELS Center in analyzing a sounding are as follows:

1. Temperature of the lower 3,000 feet. Where significant surface heating (or cooling) is expected before probable occurrence time, the actual sounding temperature is modified through this layer. In general, this modification is made on 09Z and 15Z soundings by assuming a dry adiabatic temperature lapse rate (lower 3,000 ft. only) through the predicted afternoon maximum temperature (t). Following the surface temperature forecast, the mean potential temperature ($\bar{\theta}$) of the lower 3,000 feet is determined.

2. The mean mixing ratio of the lower 3,000 feet ($\bar{w}$). This is determined graphically (equi-areal method) from the plotted sounding.

3. Lifting condensation level (LCL). The intersection of the mean mixing ratio and the mean potential temperature lines on a pseudoadiabatic chart define the LCL. Above this level, it is assumed that parcels from the lower 3,000 feet will rise along a moist adiabat without entrainment.

4. Level of free convection (LFC). Following along a moist adiabat through the LCL, the LFC is defined as the level above which the parcel is buoyed upward by its lesser density than that of its environment. This level is determined at the point where the parcel trajectory, along a moist adiabat from the LCL, crosses the actual sounding.

5. Lifted index (LI). The parcel trajectory is continued upward along a moist adiabat from the LFC to the 500-mb level. The 500-mb temperature thus determined might be considered as the parcel or updraft temperature within the thunderstorm cloud, if one develops. The algebraic difference between the environment temperature and the updraft temperature (observed temperature minus computed parcel temperature) at the 500-mb level defines the LI as used in SELS. It will be noted that the evaluation of the LI is similar to that of the Showalter stability index (SI) with the difference being in the more representative determination of the lower-level moisture and temperature values. In dealing with severe local storms, LI values are usually algebraically less than SI values.

Raob Analysis Chart.—Values from these analyses of soundings are plotted on base maps at the SELS Center, using the following station model:

| | | | |
|---|---|---|---|
| (Level of free convection) | LFC | LI | (Lifted index) |
| (Mean mixing ratio) | $\bar{w}$ | $\bar{\theta}/t$ | (Mean potential temperature)<br>(Forecast surface temperature) |

($\theta$ and t are used primarily as forecast check points after the analysis and have no further prognostic value.) Examples of the analysis of soundings are shown in figures 22 (actual sounding) and 23 (prognostic sounding). An analysis of the plotted chart is then made for values of the LFC, $\bar{w}$, and LI. Conditions favorable for the development of tornadoes include low elevations (high pressures) of the LFC, large $\bar{w}$ values, and large negative values of the LI.

For convective activity that results in severe local storms, it is reasonable to expect that the LFC would lower in time and may become coincident with the LCL at the time of thunderstorm development. One of the primary causes of a lowering LCL is a low-level injection of moisture. This acts to increase the positive area as well as to decrease the negative area. Thus the upwind gradient of $\bar{w}$, or moisture, is of considerable importance in the final forecast. Surface dew points during late morning and afternoon permit an hourly check on the change in $\bar{w}$. Local heating and convection acting on the surface layer of an air mass in which the mixing ratio decreases with height will normally decrease the surface dew point through layer mixing. Therefore, rapid or continuous increases of the surface dew point, from hourly reports, over threat areas are of particular significance since these increases represent moisture advection into these areas.

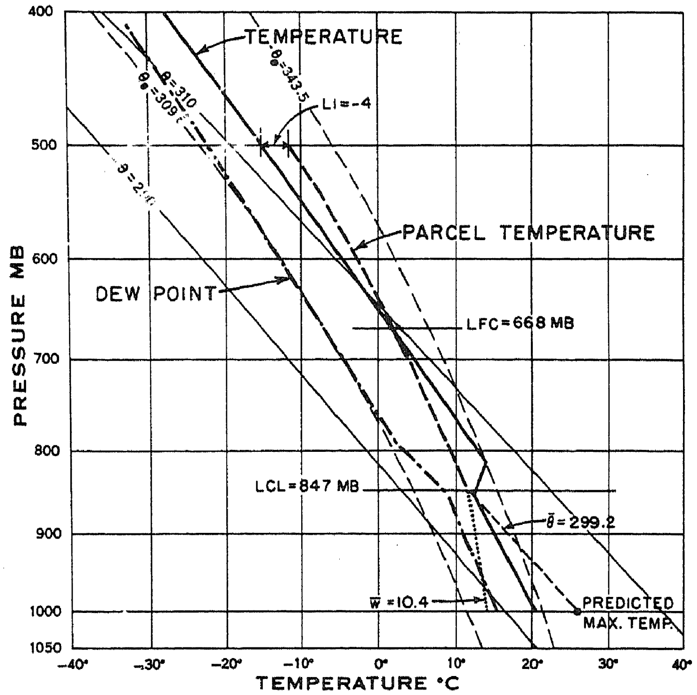

Figure 22. Example of an analysis of an observed sounding.

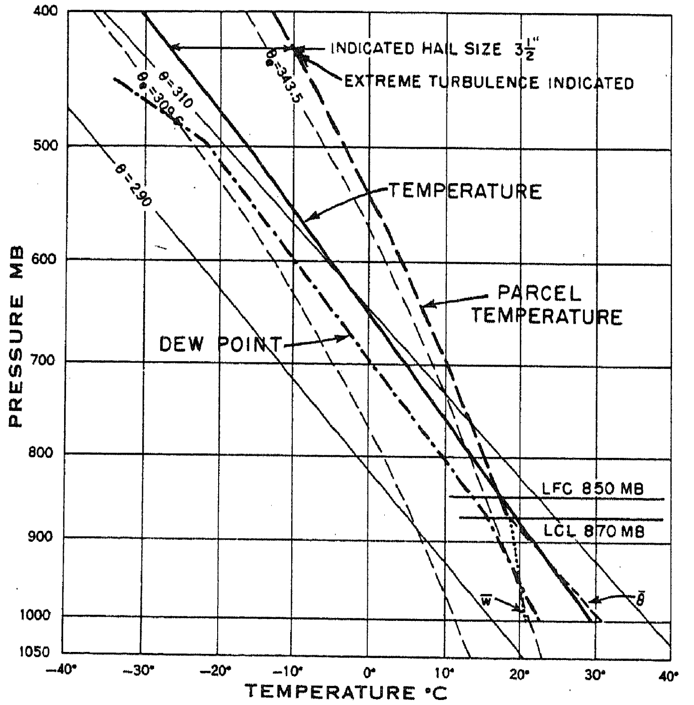

Figure 23. Example of an analysis of a prognostic sounding. Values of hail size and turbulence class are also noted.

The Prognostic Sounding.—Much of the SELS Center's routine operation is aimed at the production of a prognostic sounding which is intended to be representative of the air mass within the threat area under consideration before severe local storms develop. The prognostic sounding provides an important basis for the determination as to whether or not a severe local storm forecast will be issued. Also, if one is issued, it will provide the principal basis for forecasting hail size and class of turbulence.

The preparation of a prognostic sounding involves the prediction of temperature and humidity at selected levels above a given point at some future time. The levels involved usually include the surface, 850, 700, and 500 mb as well as the base and top of the low-level inversion when it is present. The principal processes involved in changing the temperature and humidity at a given point are as follows:

1. Temperature change.

a. The horizontal advective change. This is usually evaluated only for the 850, 700, and 500-mb levels. Isotherms are moved with the expected horizontal trajectory of air parcels, but these values are modified for the three other processes involved in temperature changes listed below. Modification usually has the same effect as reducing the advecting wind, i.e., the isotherms move more slowly than the wind (generally at about 60% of the wind speed). This is so because cold advection is usually accompanied by subsidence and heating from below, while warm advection is accompanied by upward motion and cooling from below.

b. The dynamic change. A recent paper on the release of convective instability [8] stressed the importance of the dynamic change. This change is a consequence of lifting or vertical stretching and involves adiabatic temperature changes as well as those resulting from the release of the latent heat of condensation. The significance of this change, particularly when dealing with an air mass containing a low-level inversion, may usually be estimated by simply lifting all points of a sounding. The amount of this lift varies with the situation, being greatest in the case of favorable indices of low-level convergence and high-level divergence. A lift of 50-100 mb is usually used to estimate this change but it should be understood that this is a tentative figure at this time and further study on this problem is continuing.

c. Surface heating (or cooling). The modification of the surface and low-level temperature is determined largely by the surface temperature forecast. This has already been taken into account to a certain extent in the analysis of the actual soundings. However, further consideration must be given to a surface temperature forecast for the time and site of possible occurrence.

d. The latent heat change. (Heating due to the release of the latent heat of condensation is taken into account in the preparation of prognostic soundings in the evaluation of the temperature change resulting from lifting described above.) Isobaric evaporational cooling at low levels, due to rain falling through non-saturated air, is sometimes an important factor in the determination of the temperature through the lower 3,000 feet. This effect is most frequently taken into account in the prediction of the surface temperature and also in the mean potential temperature of the lower 3,000 feet which is used to calculate the LCL. Also, there may be occasional instances in which evaporational cooling, at say 700 mb, due to rain falling from some higher level, effectively lowers the height of the LFC. The lowest temperature to which non-saturated air parcels can be cooled through evaporation alone is represented by the wet bulb temperature of the air.

2. Humidity change.

a. The horizontal advective change. An adequate low-level moisture supply is currently considered to be a necessary condition prior to the formation of severe local storms. The horizontal low-level advective moisture change is estimated from the surface map, the 850-mb chart, and plotted soundings. Moisture values are moved with the expected trajectory of air parcels and modified as indicated in the two processes listed below.

b. The vertical change. While the change in moisture values on the constant pressure charts is usually considered in most forecasting operations as resulting mostly from horizontal advection, it is important to remember that moisture also moves with a vertical component. Indeed, it is this movement that is of paramount importance in the development of convective activity. In the prognostic sounding, this vertical increase, or "pull-up", of moisture is taken into account through the same lifting process described under the dynamic temperature change above.

c. The evaporation-condensation change. The moist adiabatic lapse rate may be used to evaluate mixing ratio or dew point changes under saturated conditions during lifting or sinking. Humidity increases due to the evaporation of rain falling through non-saturated air at low levels are occasionally important in lowering the LFC and the LCL.

These steps are necessarily discussed in general terms. As in the case of the preparation of prognostic surface maps, a general outline of the factors involved can be presented whereas the relative importance of the various factors will necessarily vary from one situation to another.

### Dynamic Considerations

*Wind Field and Jet Structures.*—The importance of strong winds, or jets, in the mid-troposphere in connection with this type of forecasting has been stressed by several writers in recent years. In their first paper on tornado forecasting, Fawbush, Miller, and Starrett [19] stated "The horizontal distribution of winds aloft must exhibit a maximum of speed along a relatively narrow band at some level between 10,000 and 20,000 feet, with the maximum speed exceeding 35 knots." Beebe and Bates [8] proposed models of jet structures at the 850 and 500-mb levels, which could produce vertical stretching and thus release convective instability. A later study of some selected tornado situations by Kraft and Conner [30] found an average wind speed of the low-level (850-mb) jet of 48 knots and that the speed of the upper jet (500 mb) averaged 73 knots. Conner [13] subsequently expanded upon this work and presented 5 models of jet configurations at the 500-mb level which agree very well with those of Beebe and Bates [8].

By considering certain indications of the vorticity equation Beebe and Bates [8] have shown that there may be several configurations of jet axes and jet maxima such that low-level (850-mb) horizontal convergence is surmounted by higher level (500-mb) horizontal divergence. Vertical stretching (in the lower troposphere) and upward motion associated with these jet structures would help to release convective instability and thus be instrumental in the development of severe thunderstorms and tornadoes.

In general, the jet axis, or horizontal wind maximum, at the 500-mb level that is of concern here is that which lies to the southeast of the maximum wind center of the major axis. These secondary jets may often constitute little more than the southern or southeastern boundary of a plateau of high wind speed extending to the right (usually southeast) of the major jet axis. (An exception to this generalized case exists in the situation illustrated in fig. 27 where the 500-mb jet of interest is usually the major jet at that level.) The intensification of this secondary jet is frequently noted on 500-mb charts prior to tornado occurrence and the wind maximum at the time and site of occurrence is generally quite apparent. It is stressed that the interest in vorticity advection at the 500-mb level lies in its usefulness as an indicator of vertical motion.

The use of the following models in tornado forecasting is based primarily upon a careful analysis of jet (or wind maxima) axes at the 850 and 500-mb levels. It is assumed that, in general, the maximum winds along the jet axes occur near the intersections of the jets in these models. The prognosis of these jet axes may be accomplished through the application of contour prognoses, extrapolation, and anticipated reactions of thickness fields. The low-level jet is also closely related to surface developments and displacements.

Figure 24 is a schematic model in which an anticyclonically curved upper jet axis is oriented

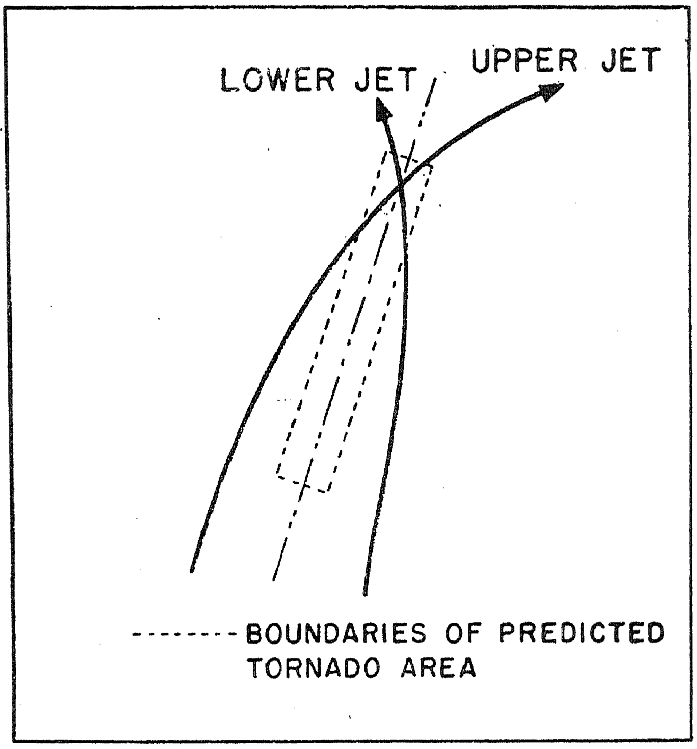

Figure 24. Model of an anticyclonically curved upper jet and cyclonically curved lower jet intersecting at an angle of less than 45°. Note the threat area outlined along the bisector line. Actually, a lower jet without curvature would usually result in about the same predicted tornado area. (From [8].)

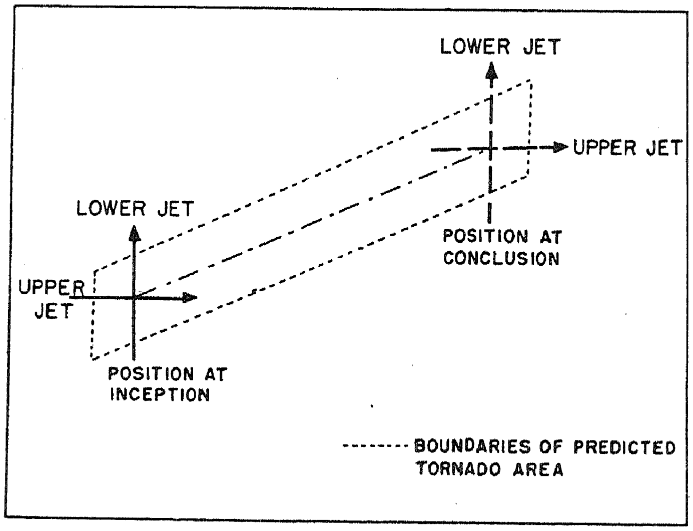

Figure 25. Model of a perpendicular intersection of upper and lower jets without curvature showing predicted tornado area as jet intersection moves in the indicated direction. (From [8].)

at an angle of 45 degrees or less with respect to a lower jet axis. With this structure, the significant area is related to a line from the point of intersection (or the extended intersection point if the jets do not actually intersect) southwestward, bisecting the angle between the two jets. The probability of such interaction is a function of the spatial separation of the jets and, thus, of the angle between them. Occurrences usually break out first in the south and move northward along the line bisector. This pattern is frequently quasi-stationary and attends many major family outbreaks. A typical predicted tornado area is outlined in figure 24 (no movement is shown).

Figure 25 is a model of a right angle intersection of jets without curvature. A configuration of jet axes is predicted for a given point and this intersection is further predicted to move across the field shown. Since the fine structure of the jet is not amenable to treatment, the area chosen for the prediction of tornado occurrence is centered upon the line of movement of the jet intersection and embodies an area of uncertainty, due to lack of detailed data, on either side of this line. The occurrences are usually found very close to the intersection and seldom consist of more than one or two tornadoes.

Another model (fig. 26) is that of an anticyclonically curved jet axis aloft which intersects the low-level jet at an angle of more than 45 degrees. The reasoning and the extrapolation of the intersection point are similar to that in the case above (fig. 25). The principal difference is the weighting of the area to the southwest of the upper jet maximum and to the west of the lower jet axis for the probability of occurrence. The limits of error in positioning the jet axes require inclusion of the intersection point and some area to the north and east. A typical area for tornadoes with such a configuration is outlined in figure 26.

A fourth case, figure 27, is that of the intersection of a cyclonically curved jet aloft with a low-level jet. This situation is also similar to the second example but weight is given to the area to the northeast of the upper jet maximum. (The differences in weighting of areas in figure 24-27 are explained in the original paper [8].) This case is representative of the "cold" type of occurrence, such as the storm at Hartford, Conn., May 19, 1954. In these situations, the low-level jet is not necessarily well defined, but surface indices place the occurrence in a region of pronounced low-level convergence.

The choice of the 850 and 500-mb levels for these indications of convergence-divergence was mainly a matter of convenience since these charts are analyzed on a routine basis. Also, wind reports at 500 mb have generally been more numerous and accurate than those at higher levels, However, there are many occasions when these levels are not sufficiently representative of the strong wind patterns at low and high levels.

Thus, at low levels, the gradient-level wind (defined as the strongest wind reported within the first 4000 ft above the station) is plotted on sectional surface charts within and surrounding threat areas. The low-level jet or other convergence indices may then be determined from streamline and isotach analyses of these winds in conjunction with surface map indications such as isobaric gradients or isallobaric fields.

At higher levels it has been found that a single isobaric surface (such as 500 mb) does not satisfactorily portray the location of the jet stream(s). Therefore, procedure at the SELS Center now includes the preparation of a "Jet Chart", upon which are plotted the direction, speed, and height of the strongest wind in the 20,000 to 45,000-ft layer for each wind reporting station. Streamlines and isotachs are drawn on this chart utilizing thermal and contour gradients at the 500, 400, 300, and 200-mb and tropopause levels. From the analyzed wind field on this jet chart vorticity and divergence fields can be estimated quickly in qualitative fashion. Short-period prognoses (up to 12 hours) of the wind field may be made with reasonable accuracy using methods described in Appendixes 5 and 6 of [40]. The result is that the path of high-level divergence areas which may be favorable to severe storm development may be predicted with reasonable accuracy for a period up to 12 hours in advance. Experience has shown that these jet stream considerations are easily applied and are useful in most severe storm forecasts.

Jet configurations, of course, constitute only one means of effecting vertical stretching, or lifting, of an air mass. Other mechanisms such as cold fronts, warm fronts, upper lows or troughs without pronounced jet structures, terrain features, etc., exist and must be given consideration. Thus, the lack of prior existence of these jet structures does not preclude the expectancy of severe local storms, i.e., a low-level jet may not be in evidence but may result from marked deepening of a low center. Particularly in summer, the high-level jet may not be indicated by winds aloft measurements but will develop subsequently as a result of low-level convergence and increasing thermal gradient.

More exact quantitative methods for determining the divergence field and predicting it for short periods should produce better results than the present qualitative estimates. It is hoped that some use may be made of vertical motions that are calculated numerically by the Joint Numerical Weather Prediction Unit in Washington. The problem of obtaining accurate, quantitative estimates of divergence fields is now under investigation in the SELS Center.

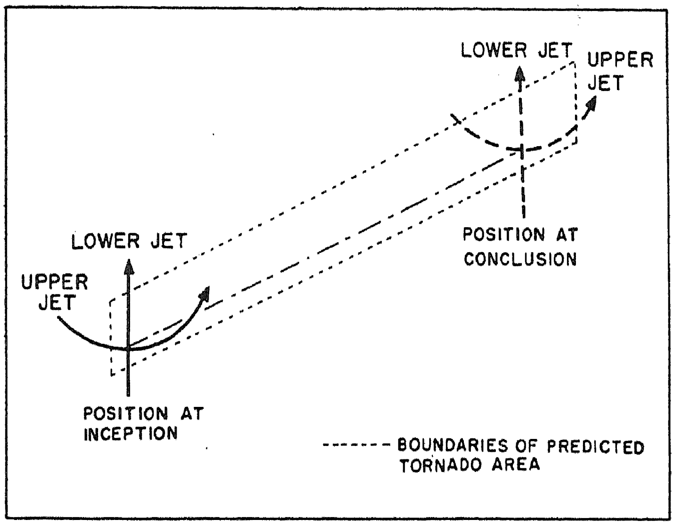

Figure 26. Model of an anticyclonically curved upper jet intersecting a lower jet, without curvature, at an angle of 45° or more with predicted tornado area as jet intersection moves in the indicated direction. (From [8].)

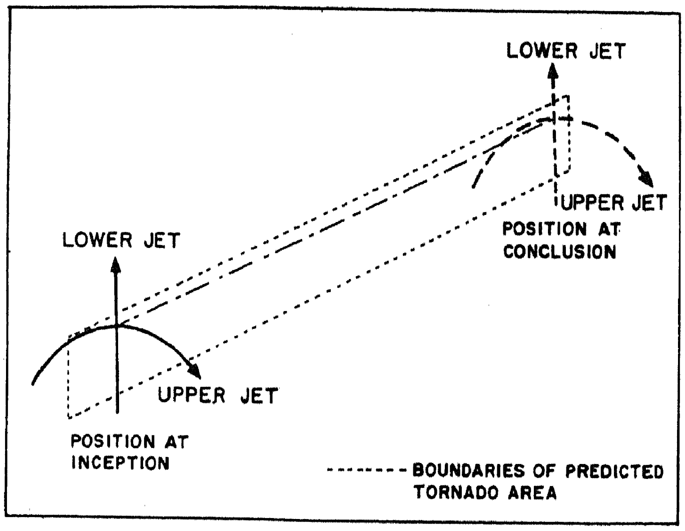

Figure 27. Model of a cyclonically curved upper jet intersecting a lower jet without curvature showing predicted tornado area as jet intersection moves in the indicated direction. (From [8].)

The 700-mb No-Change Line.—The no-change line is defined as a line connecting points of no advective temperature change on constant pressure charts – most often at the 700-mb level. This line separates areas of warm advection from those where cold air advection is indicated at sounding time. There may be, of course, several no-change lines on a 700-mb chart which covers the continental United States, but in the forecasting of severe local storms the 700-mb no-change line of primary interest is the one just ahead of the trough line. In general this no-change line will be observed near the isothermal ridge at this level. While the use of the no-change line is frequently believed to be important in this type of forecasting, there are many instances when this is not applicable, particularly in cases of flat or indefinite gradients.

The use of the no-change line in forecasting has been discussed by Van Thullenar [54] and others, but in the forecasting of severe local storms it serves two functions. When it separates areas of strong cold air advection from pronounced warm air advection, it calls the forecaster's attention to developmental situations. It has been observed that the advancement of the leading edge of the cold air advection (i.e., the no-change line) into a position ahead of a pronounced trough is usually associated with deepening or intensification of surface low centers.

A second function of the 700-mb no-change line is to delineate the eastern edge of an area where the air mass is undergoing a change from one of increasing stability (warm advection) to one of increasing instability (cold advection) in the mid-troposphere. It is believed that much of the cooling observed at this level is dynamically induced so that the movement of the no-change line may be independent of wind speed at that level. Even though the magnitude of apparent cold air advection does not often appear to be strong enough in itself to cause the intense instability at the time and site of severe local storms, it seems at least to assist in this process in most cases. Several writers [11, 14, 23, 33] have attributed much of the intense instability of such storms to cold air advection at some level above a layer of warm, moist air near the surface. However, experience thus far in the forecasting of these storms seems to largely discount the importance of such horizontal temperature advection as the cause of the intense instability.

It will be noted that there may appear to be an inconsistency between neutral or slight cold advection at 700 mb, as indicated by the no-change line, and warm advection in the 850-500 mb layer, which is implied from the wind shears of the jet models shown in the preceding section. Observations have shown that the usual tornado situation is one in which there is marked warm advection in the layer from the gradient level to near 700 mb, while slight cold advection is usually in evidence from near this level upward through 500 mb. Thus, the shear from the gradient (or 850-mb) level to the 500-mb level may show net warm advection, but most of this occurs in the lower layers in these severe storm cases.

Pressure Jumps.—The pressure jump concept provides a physical model for the mechanism of squall lines [46,47,48]. Considerable data have been compiled to show the association between pressure jumps and tornadoes. In a recent study, Tepper [49] found that 68 percent of those tornadoes whose time and location were known (65 tornadoes) corresponded within one hour to the pressure jump line isochrone. In this same study it was found that most of the pressure jump lines had a life of 4 hours or less and that most of the severe local storms occurred during the first 2 hours of the life of a pressure jump line. Thus the time lag between the development of a pressure jump line and tornado formation is usually too short to be utilized in current forecasting and distribution methods. Indeed, as Newstein [37] pointed out in a recent paper, a tornado occurrence may precede the formation of a pressure jump line.

Pressure jumps are often associated with intense convective storms so that the times and locations of pressure jumps, along with other indicators of severe local storms, are carefully followed on activity charts. Pressure jump reports may often be used as an index to the intensity of thunderstorms and, therefore, as an analysis tool in tracking the movement of instability lines after they have formed. However, forecasters who must deal with the problem of predicting areas favorable for tornado occurrence must keep in mind the fact that many pressure jumps occur without tornadoes and, at least with the present synoptic reporting network, it appears that tornadoes may occur without prior evidence of pressure jumps.

## Formulation of the Tornado Forecast

In common with other types of weather forecasting, the formulation of a tornado forecast does not entail a simple weighing of parameters or direct application of rules. Instead the forecaster must arrive at some decision, or series of decisions, in each individual situation regarding the combined relative importance of the climatological, synoptic, thermodynamic, and dynamic factors previously discussed. So, until these factors, as well as others yet to be developed, can be measured and combined in a more objective manner, tornado forecasting must rely heavily upon judicious, subjective evaluation of each particular severe storm situation, real or potential.

The development of a set of rules which would be applicable in any area, any season, or any of the four air mass types previously described has not been attempted. Nevertheless, the following discussion will attempt to portray the general manner in which the methods previously discussed in some detail are brought together in the formulation of a tornado forecast.

The formulation of tornado forecasts is a function of dynamic, thermodynamic, and synoptic-climatological considerations. There are occasional instances in which all of the thermodynamic considerations are pronounced and, at the same time, all dynamic considerations are strong and well-marked. In such cases (e.g., the tornado situation of March 21-22, 1952), nearly all criteria for forecasting tornadoes are fulfilled. However, these situations are not representative of those which most frequently confront the forecaster. Experience has shown that, in general, dynamic considerations are the strongest during late fall, winter, and spring when thermodynamic factors may be weak. These are, as is well known, the seasons when the more severe outbreaks occur. On the other hand, thermodynamic considerations seem to have a more dominant role during summer and early fall when pressure patterns are usually weaker and less well-defined.

A first approximation to the section of the United States in which tornadoes are a threat can be made from the following considerations:

1. Stability index. Areas of negative values, particularly large negative values, become tentative suspect regions for further analysis.

2. A quick estimate is made of the conditions discussed under "Synoptic Considerations."

3. Climatology. The above gross considerations thus outline a region, perhaps as large as several States, for further refinement by the following:

A. Synoptic considerations.

a. 6 and 12-hour prognostic surface charts, including fronts, instability lines, and pressure centers.

b. Deepening. Deepening surface lows, or waves, are known to be favorable for tornado formation, particularly of the family-type outbreaks in winter, spring, and early summer. It should be noted that there are also many tornado occurrences without prior evidence of surface deepening.

B. Dynamic considerations.

a. 6 and 12-hour prognostic positions of 850 and 500-mb jets. The jet structures, as previously described should, in general, be in evidence or predicted over the threat area.

b. 700-mb no-change line. In the usual situation during winter, spring, and early summer, it should be expected that the 700-mb no-change line will be coincident or nearly coincident with the 850-mb warm tongue at the time of instability line development.

These considerations will usually further limit the tentative region to a threat area in which tornadoes may be possible but not necessarily probable.

C. Thermodynamic considerations.

a. Initial analyses of observed soundings should in general show LFC values that are below the 600-mb level (i.e., at lower elevations, or higher pressure) within threat areas. Analyses of prognostic soundings should show LFC's around the 850-mb level. Other things being equal, the expectancy of severe local storms is directly proportional to higher pressure values (lower elevations above the ground) of the LFC. Both convective and conditional instability should exist, or be predicted, near the time and site of tornado occurrence.

b. LI values, within a threat area, should be negative and the probability of severe local storms is proportional to increasingly large negative values of the LI. In most cases, large negative LI values and low elevations (high pressure) of the LFC indicate that there are large positive areas associated with the soundings.

c. Moist layer. In general, the moist layer should be at least 3,000 feet deep, but in cases where the minimum is only slightly exceeded there must be evidence of a strong moisture injection within this layer to replenish the supply if and when convective activity occurs. Those cases of moist layer depths of 3,000-8,000 ft that suddenly increase to above the 700-mb level are believed to constitute direct observational evidence that convective instability is either being released at sounding time, or will be released very soon. Therefore, such areas are "hot" and quick forecast action is advisable. In general, SELS forecasters attach little significance, per se, to the existence or non-existence of a relatively dry layer overlying the lower moist layer, except as an index to convective instability.

4. The prognostic sounding(s) is (are) intended to represent the air mass structure within the threat area(s). Thus, it constitutes the final thermodynamic consideration as to whether or not a tornado forecast will be issued. Also, computations of hail size and extreme turbulence are made from this sounding.

An exception to these considerations has been noted in the case of the jet structures shown in figure 27. In these cases there may be moisture to great heights over a very large area and static SI values may be positive (but becoming negative at the time and site of the beginning of activity).

Also, surface dew points may be in the 40's. In these cases, the instability is assumed to result largely through dynamic modifications of the air mass, although high-level cold advection may also be important.

## EXAMPLE OF A TORNADO FORECAST

(Excerpted from an article by Lee [31].)

One of the most death-dealing series of convective storms of the last few years occurred during the afternoon and evening of February 1, 1955 - the first day of the 1955 "tornado season." ...In many respects, this situation presents nearly ideal examples of parameters used in forecasting tornadoes. At the surface, a pronounced Low was moving to the east-northeast at a moderate rate; an mP cold front followed in a V-shaped trough extending south-southwestward; warm, moist air with dewpoints in the 50's was moved northward ahead of the cold front by strong southerly gradient winds; and marked pressure falls occurred ahead of the Low. At the 850-mb. level, a sharp moisture injection, nearly coincident with the temperature ridge [14], was associated with a band of strong winds at this level. One feature of this situation that is different from many tornado outbreaks was the shift in winds at this level from a strong south-southwesterly straight flow at 2100 CST, January 31 to a strong cyclonically-curved flow by 0900 CST February 1. One of the most significant features at the 700-mb. level was the appearance of a moisture "bubble" at Little Rock, Ark., at 0900 CST. At the 500-mb. level, a strong cyclonically-curved jet moved into the area in a nearly ideal manner.

Features at the surface. The surface charts prior to the development of these storms indicate that the associated surface cold front was the leading edge of a mass of Pacific air that entered the west coast early on January 30. The southern portion of this front moved eastward at 20-25 knots to merge with a 1012-mb. low center in central Arizona. The front accelerated to 25-30 knots on January 31 and by noon on February 1 the Low had deepened to 1003 mb. in the vicinity of Springfield, Mo. Positions of the surface fronts and pressure centers prior to the occurrence of the tornadoes are shown in figure 28.

Hourly surface charts which were drawn from 0830 CST through 2030 CST, February 1, are not reproduced here because the 3-hourly maps shown in figure 29 are sufficient for continuity. (Analyses shown here were taken from the Severe Local Storms Center working charts.) The squall line and weak secondary warm front at 0930 CST (fig. 29A) were being followed...for indications of possible severe storm development. One interesting feature of this series of surface maps is that the relative positions of the surface dew point ridge and the squall line remained nearly coincident throughout the life of the squall line, even in no-precipitation areas.

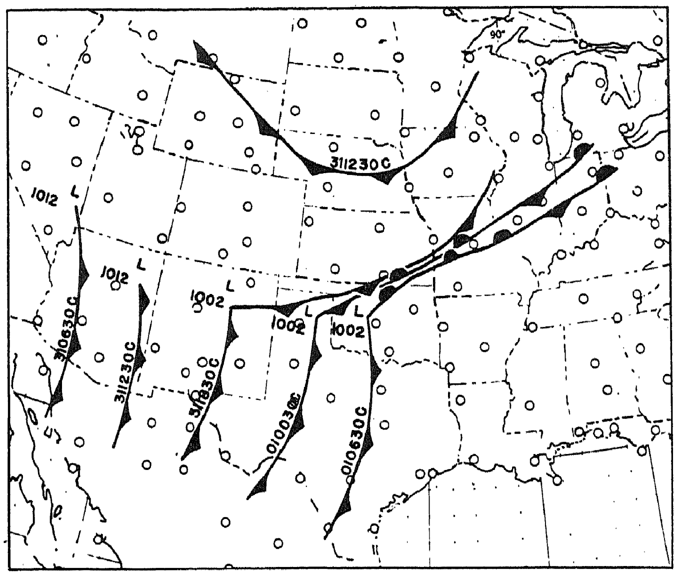

Figure 28. Times and positions of surface fronts and low centers prior to development of storms, 0630 CST, January 31 to 0630 CST, February 1, 1955.

The low center near Tulsa, Okla. at 0930 CST moved eastward at 45 knots between 0930 and 1230 CST while the cold front and squall line also moved eastward at 45 knots. Around 1230 CST (fig. 29B), the cold front apparently decelerated to 30 knots, and weather activity along the front seemed to decrease, while activity along the squall line began to increase. A micro-low at 1330 CST was detected at the intersection of the squall line and secondary warm front. The echo of a rain area associated with this micro-low was picked up by radar at both Little Rock, Ark. and Memphis, Tenn. As the squall line moved eastward, this echo moved to the northeast and expanded into a line of echoes 70 miles long and 5 miles wide with an isolated echo about 20 miles farther south. When the line of echoes passed over Memphis at 1440 CST, its northern edge was some 25 miles north of Memphis. The storm damage in Arkansas occurred with the passage of the squall line and at its intersection with the secondary warm front. At 1530 CST (fig. 29C), the squall line and its associated severe storms were advancing eastward at 45-50 knots. The micro-low continued to be in evidence and a ridge of high pressure formed between the squall line and the cold front. While the central and northern portions of the squall line were active during the afternoon and evening, the severe damage was confined to the

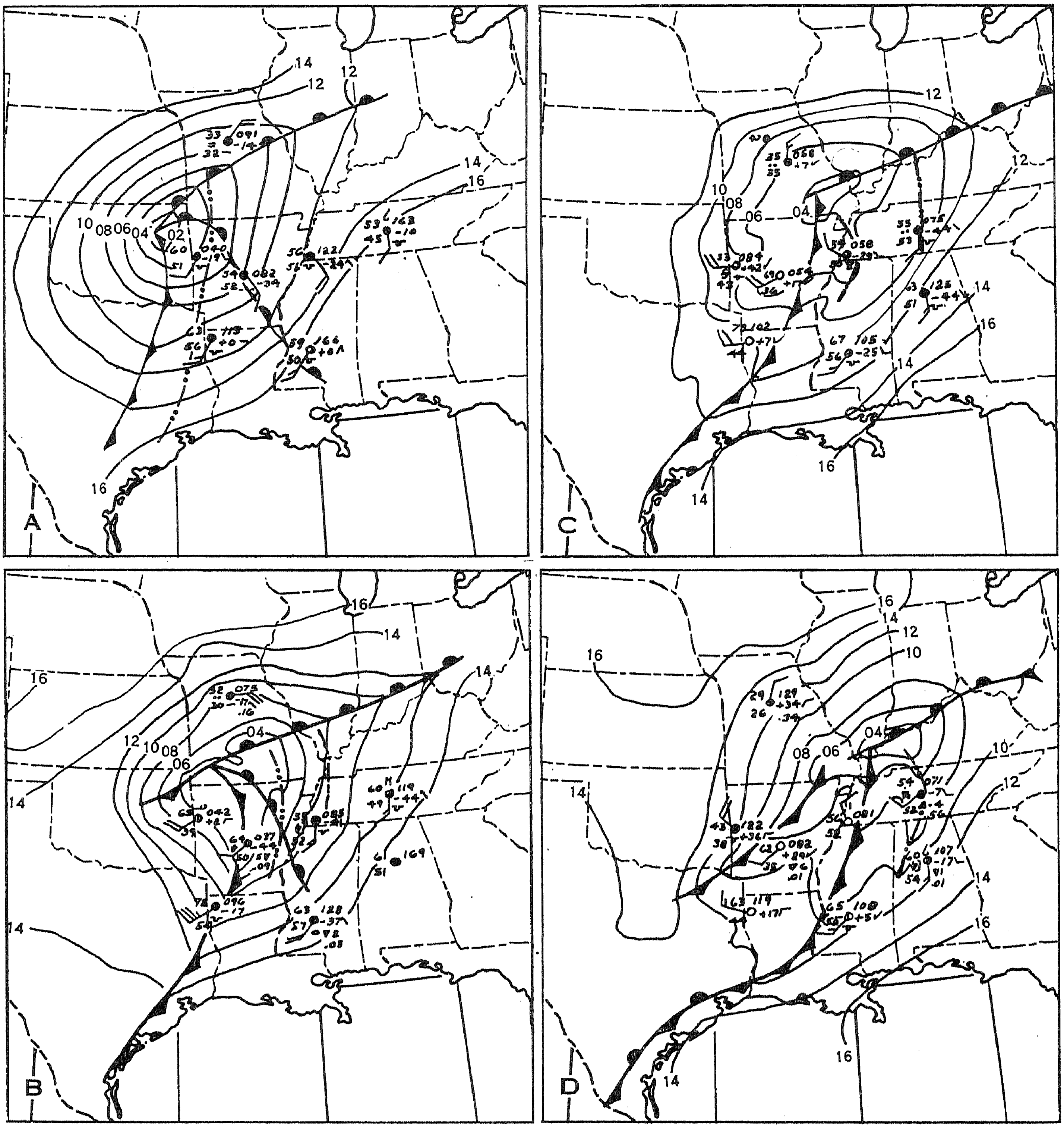

Figure 29. Three-hourly sectional surface maps, February 1, 1955, for (A) 0930 CST, (B) 1230 CST, (C) 1530 CST, and (D) 1830 CST. Analyses are from the Severe Local Storms Forecast Center working charts.

vicinity of the micro-low which continued eastward to the Huntsville, Ala. area by 1830 CST. The center of 3-hourly pressure falls (-4.0 mb.) moved from central Arkansas at 0930 CST to northeastern Alabama (-5.0 mb.) at 1530 CST. These large pressure falls were accompanied by large pressure rises to the west; +6.0 mb. in north central Texas at 0930 CST and +6.0 mb. near Tulsa, Okla. at 1530 CST.

Features at 850 mb. The positions of the 850-mb. temperature and moisture ridges at various times prior to storm development are shown in

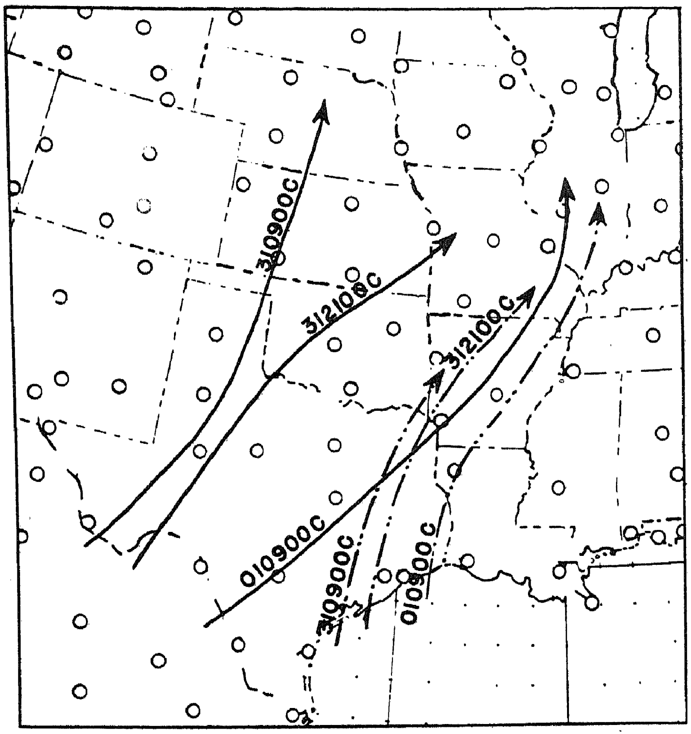

Figure 30. Times and positions of 850-mb temperature ridge (solid lines) and 850-mb moisture ridge (dashed lines) prior to development of storms.

figure 30. At 0900 CST, January 31, a warm ridge extended from west Texas to central Nebraska and a moisture ridge was indicated about 300 miles to the east. During the next 24 hours, the warm ridge moved eastward at about 17 knots while the axis of the moisture ridge moved eastward at about 5 knots. At 0900 CST, February 1, the temperature and moisture ridges were approaching coincidence. It is interesting to note that an extrapolation of these positions and movements indicates coincidence near Memphis around 1500 CST. Figures 31A, B, and C show the 850-mb. features at 0900, 1500, and 2100 CST, February 1.

An important change in the flow at this level occurred between the charts for 2100 CST January 31, and 0900 CST February 1. The earlier chart (and even 12 hours prior to that) showed a strong moisture injection resulting from a strong south-southwesterly flow from the Gulf of Mexico. However, at 0900 CST, February 1 (fig. 31A) the winds over Arkansas intensified along with the development of marked cyclonic curvature. Such a cyclonic curvature of the flow at this level tends to cut off the northward flow of moisture and is not normally associated with the formation of tornadoes.

Features at 700 mb. At 700 mb., the line of "no-change", separating the area of apparent warm

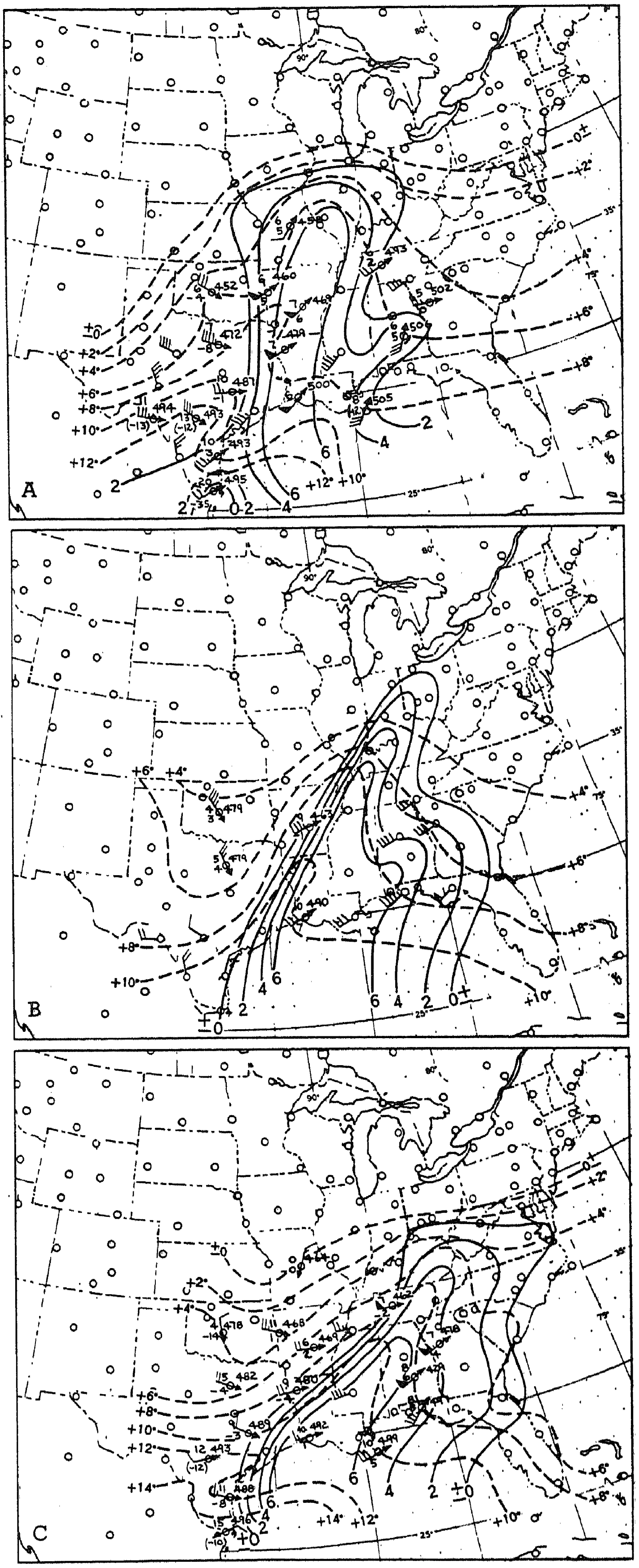

Figure 31. 850-mb charts for February 1, 1955, showing isotherms (dashed) and dew point lines (solid) for 2 °C intervals at (A) 0900 CST, (B) 1500 CST, and (C) 2100 CST.

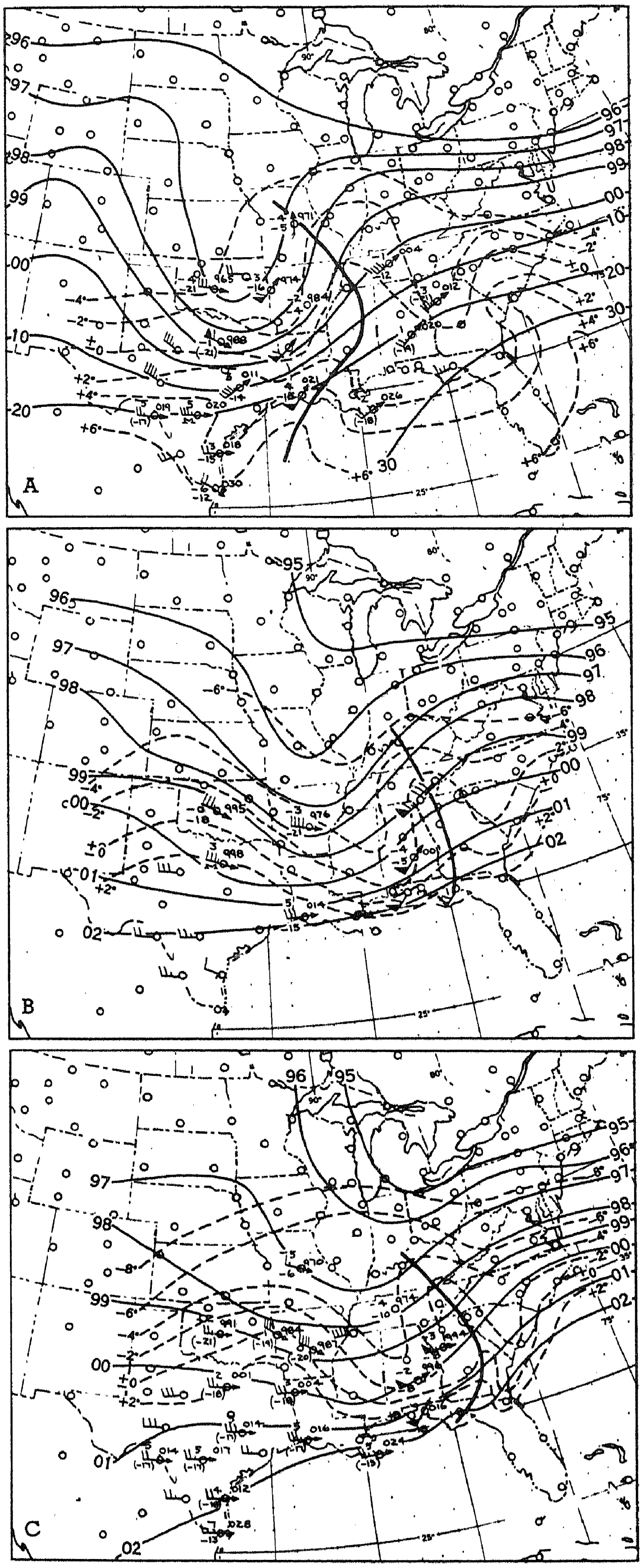

Figure 32. 700-mb charts for February 1, 1955, showing contours (light solid lines) for 100-ft intervals, isotherms (dashed lines) for 2 °C intervals, and the "no-change" line (heavy solid line) at (A) 0900 CST, (B) 1500 CST, and (C) 2100 CST.

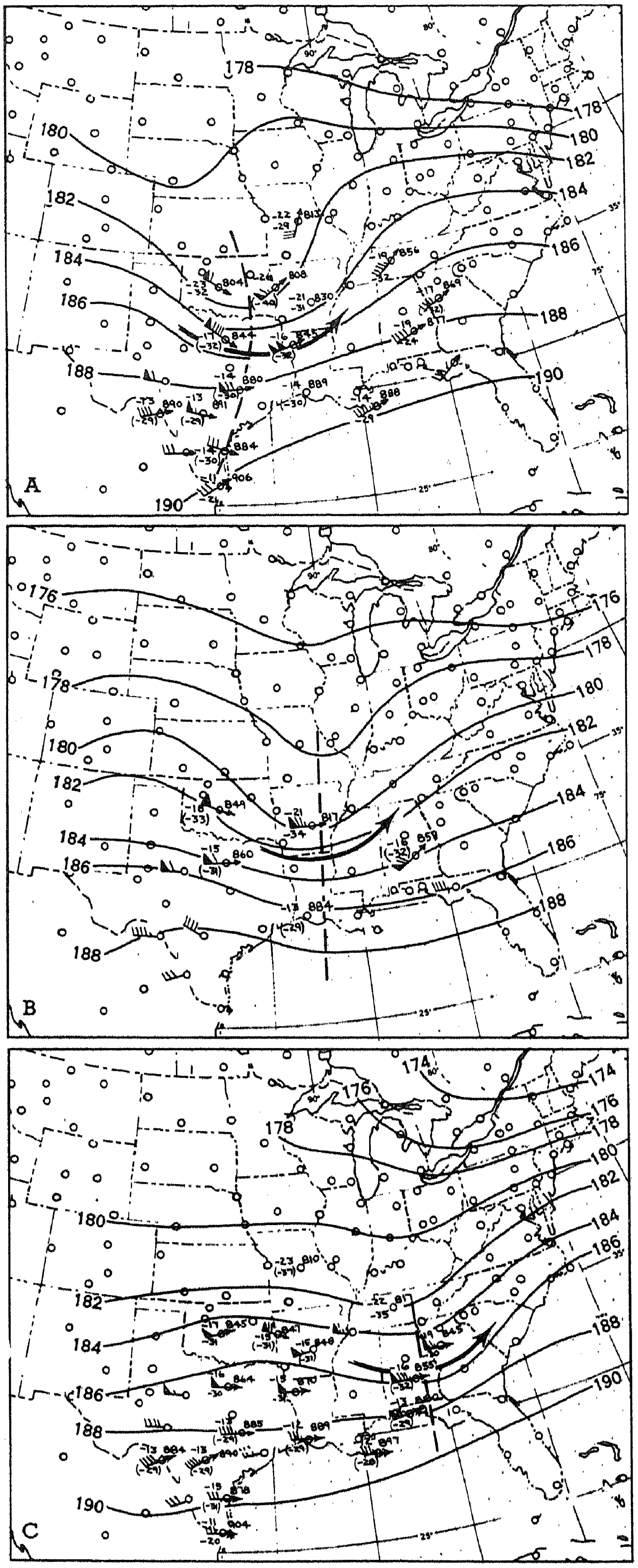

Figure 33. 500-mb charts for February 1, 1955, showing contours (solid lines) for 200-ft intervals, trough line (dashed) and the current jet position (curved arrow) at (A) 0900 CST, (B) 1500 CST, and (C) 2100 CST.

air advection to the east from the following cold air advection to the west progressed eastward and was generally followed by a 4° C. drop in temperature which tended to decrease the stability of the air mass. The positions of this line at 0900, 1500, and 2100 CST, February 1, are shown in figure 32.

As previously mentioned, one of the significant features at 700 mb. was the "bubble" of moisture that suddenly appeared over Little Rock as indicated by data at 0900 CST, February 1. Such a high moisture value (dew point of -4°C.) could not have resulted from horizontal advection since prior upstream values were much lower. Thus it must have been a consequence of an increase in depth of the moist layer through vertical motions rather than horizontal advection.

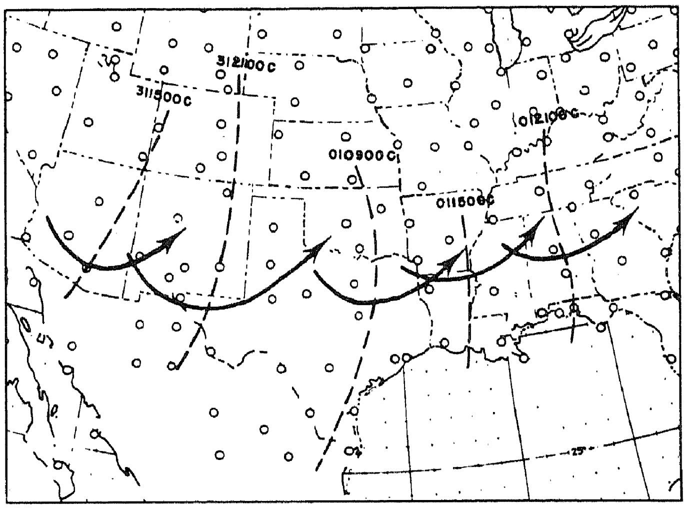

Figure 34. Times and positions of the 500-mb trough (dashed line) and jet (heavy arrow) prior to and during the development of storms, 1500 CST January 31 to 2100 CST February 1, 1955.

Features at 500 mb. The 500 mb. charts for 0900, 1500, and 2100 CST, February 1, are shown in figure 33. The times and positions of the 500-mb. jet and trough lines, taken from the Severe Local Storms Center working charts, are shown in figure 34. According to the model of jet structures proposed by Beebe and Bates [8], the left front quadrant of a 500-mb. jet axis maximum with cyclonic curvature, being an area of divergence at this level, is a contributing factor in effecting the release of convective instability. The jet positions shown at both 0900 and 1500 CST would require positioning a potential severe local storms forecast area to the left (north) of the indicated 500-mb. jet.

Upper air conditions. The Little Rock upper air soundings showed the depth of the moist layer to be near 5,000 feet at both 0900 CST (fig. 35A) and 2100 CST (fig. 35B) on January 31. However, a marked change occurred during the next 12 hours and by 0900 CST February 1 (fig. 35C), the moisture had penetrated to a much greater depth, at least 13,000 feet. At Barksdale Air Force Base, Shreveport, La., a similar situation was noted in the soundings. During the 6-hour period from 2100, CST, January 31 (fig. 36A) to 0300 CST, February 1 (fig. 36B), the depth of the moist layer remained near 5,000 feet. During the next 6-hour period, the moisture depth increased to near 8,000 feet (fig. 36C). If the base of the Barksdale inversion noted at 0300 CST were lifted 80 mb. to the 770-mb. level, the same temperature would be obtained as that observed. Further study of the changes in the lower levels of the air mass represented by

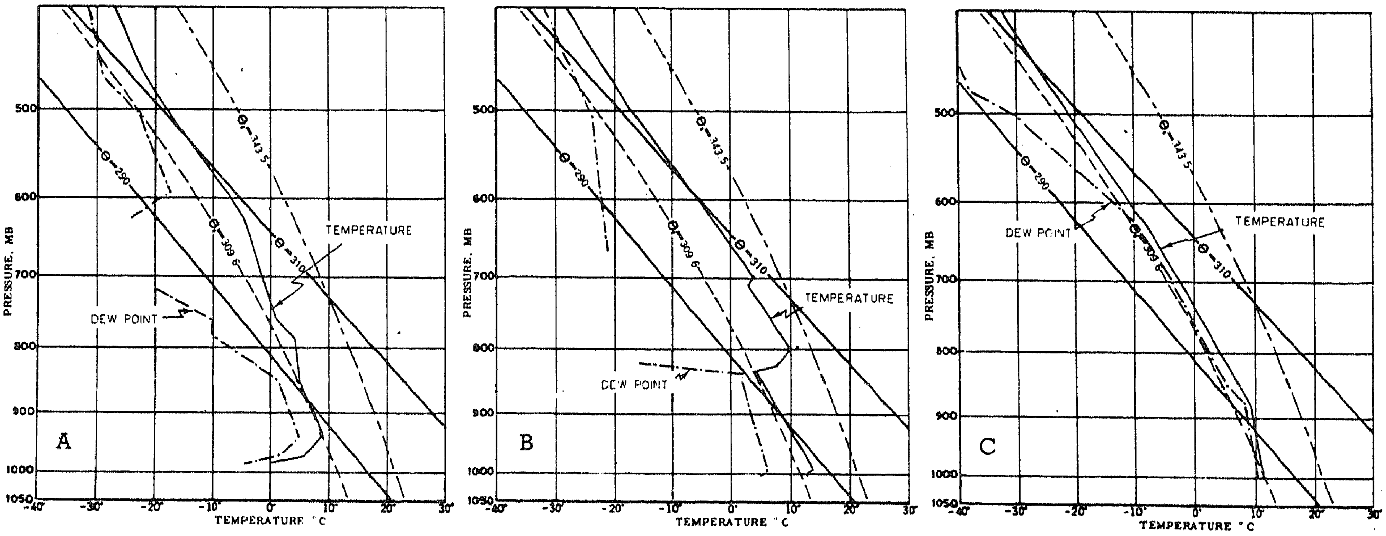

Figure 35. Upper air soundings at Little Rock, Ark. at (A) 0900 CST, January 31, (B) 2100 CST, January 31 and (C) 0900 CST, February 1, 1955.

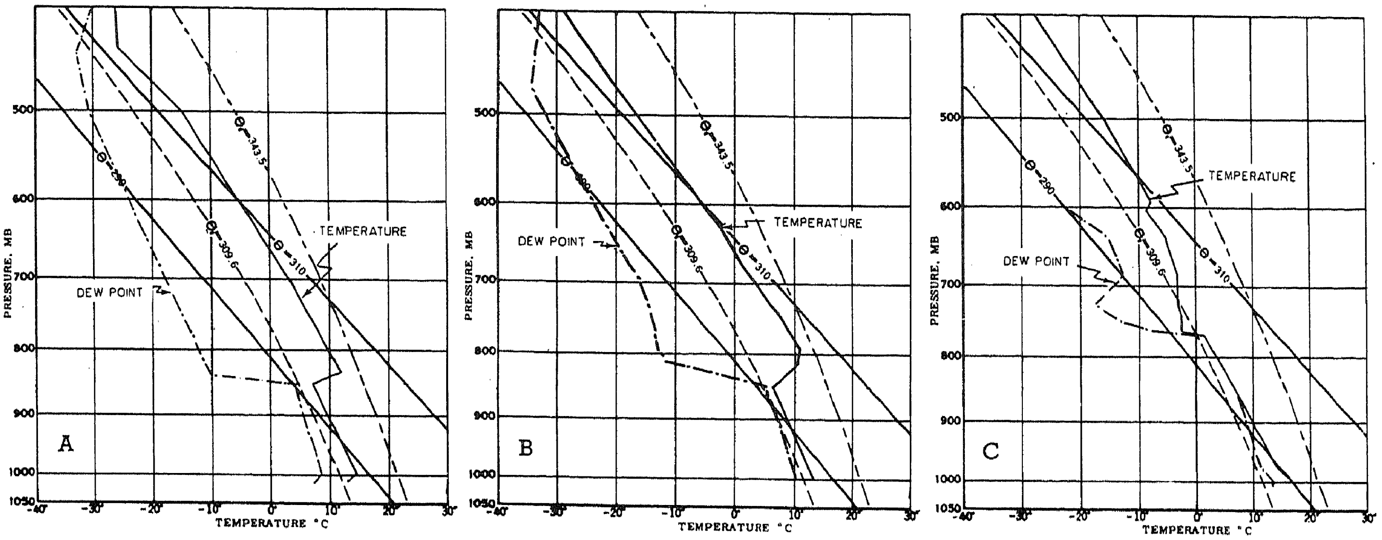

Figure 36. Upper air soundings at Barksdale Air Force Base, Shreveport, La., at (A) 2100 CST, January 31, (B) 0300 CST, February 1, and (C) 0900 CST, February 1, 1955.

the difference of the Barksdale sounding at 0900 CST from that at 0300 CST indicate that this change can be explained through a lift of 80 mb. Considering the moist layer of the Little Rock sounding at 2100 CST, January 31, and increasing the mean temperature by 2° C. and the mixing ratio by 1 gram [per kgm], the sounding observed at 0900 CST, February 1 could be obtained by lifting the lower 5,000 feet by 150 mb.

One interesting feature of the 0900 CST soundings at Barksdale, Maxwell Air Force Base at Montgomery, Ala.; Burrwood, La. and Atlanta, Ga., (all within the warm air mass), was the presence of a superadiabatic lapse rate at the top of the moist layer... At this time, the moist layer was around 7,500 feet deep except at Burrwood where the depth was near 5,000 feet. The relative humidity at the base of these superadiabatic lapse rates varied from 90 to 100 percent at Barksdale and Maxwell to around 70 percent at Atlanta and Burrwood. Precipitation was not observed at these stations just prior to or during these soundings.

Forecasting tornado development. The forecasting of this tornado development by the Severe Local Storms Center was based largely upon (1) the recognition of the strong dynamic factors evidenced on the earlier charts (strong 850- and 500-mb jet structures and intensification of the surface Low over Oklahoma); (2) a prognostic sounding for northeastern Arkansas which took into account a lift of 100 mb; and (3) the confirmation at 0900 CST by both the Barksdale and Little Rock soundings that the lower layer was undergoing lifting. Actually, the 0900 CST Little Rock sounding evidenced a greater lift than 100 mb. (at least 150 mb), while the Barksdale sounding showed less than

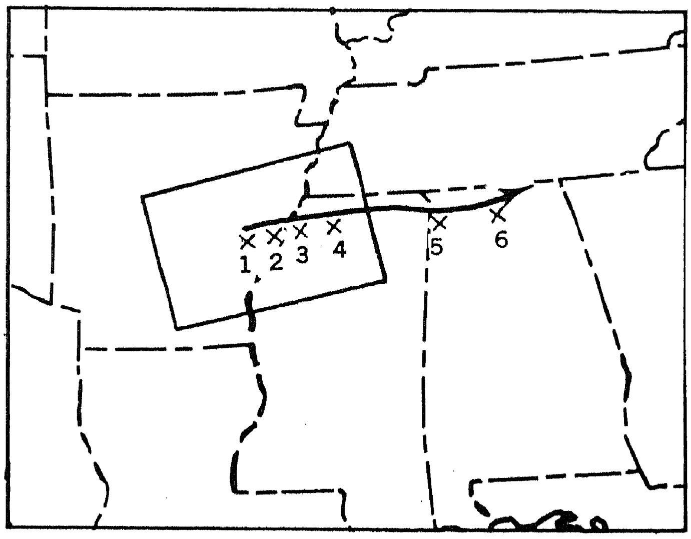

Figure 37. Location and times of reported storms and Severe Local Storms Center tornado forecast area (outlined) on February 1, 1955. Numbered x's signify:

1. 1300 CST, near Stuttgart, Ark. Initial radar echo of squall line.
2. 1400 CST, near Marianna, Ark. Minor damage and some injuries.
3. 1420-1430 CST, Robinsonville and Commerce Landing, Miss. Extensive destruction, several deaths, and numerous injuries.
4. 1440-1445 CST, Lewisburg, Miss. Extensive destruction, several deaths, and numerous injuries.
5. 1700 CST, Maud, Barton, and Mynot, Ala. 20 houses severely damaged and several injuries.
6. 1800-1830 CST, near Huntsville, Ala. Minor damage and some injuries.

100-mb. lift, so that the lifting mechanism was even more potent than expected from earlier data. Also, these considerations pointed to a threat area to the east of Little Rock. Static instability indices at 2100 CST, January 31, were all positive but by 0900 CST, February 1 the Showalter stability index...lowered to -1 at Little Rock and to -5 at Ft. Smith, Ark. The lifted index...on the prognostic sounding for 1500 CST was -4 with a level of free convection around 880 mb. While only a moderate degree of instability was available, the indications of a very strong dynamic modification of the air mass tipped the balance in favor of a forecast for tornadoes rather than severe thunderstorms.

The area was determined largely through the positions expected to be swept out by both the divergent quadrant (left, front) of the jet maximum at 500 mb. and the convergent quadrant (left, rear) of the jet maximum at 850 mb. These considerations, plus the Little Rock sounding and prognostic positions of the surface Low, located an area to the left (north) of both the 850- and 500-mb. jets. In such cases the cut-off between violent convective activity and little activity would be expected to be sharply drawn at, and to the right (south) of, the upper jet axis. The forecast area and reported storms are shown in figure 37.

# PART II–RELATED SEVERE THUNDERSTORM FORECASTING PROBLEMS

## HAIL SIZE

At the present time, hail size forecasts are of more value to aviation than other interests. Hail stones of a size smaller than 3/4 inch do not often cause important damage to commercial aircraft in operation today. So, in predicting severe thunderstorms, SELS forecasters are concerned with hail size only when hail stones having a diameter of 3/4 inch or larger are expected. At present, no attempt is made to pinpoint the exact time and location of large hail occurrences. As in the case of tornado forecasting, a prediction is made of large hail occurrence somewhere over a given area, say 100 by 200 miles, during a specified time period. Such a prediction does not carry any implication regarding the areal extent of the hail, large or small, which might be damaging to other interests, such as agriculture.

A technique for predicting hailstone size has recently been developed at the SELS Center [22] and is in current use. The mechanics of this method are quite similar to the empirical procedure of the Air Force Severe Weather Warning Center [16]. The SELS method was derived under the following assumptions: a) hail size is proportional to the updraft velocity within a thunderstorm; b) the updraft velocity results from parcel buoyancy; and hence, c) the updraft may be approximated from positive area measurements on a thermodynamic diagram. The use of this technique is based upon forecasts of severe thunderstorms made in conjunction with considerations other than positive areas or thermodynamic indices alone.

Entrainment has been neglected here. In the case of the intense thunderstorms which produce large hail sizes, the parcels may be affected but little by entrainment at the core of strong updrafts of large cross-section where the larger stones are visualized to form. However, the spectrum of hail sizes observed over an area may be due to a spectrum of entrainment rates in a complex of thunderstorm cells. Undoubtedly, entrainment must be responsible for a considerable number of failures experienced with any hail forecasting technique which does not take its effect into account (and none in application do). In general, it should be expected that the effect of entrainment in reducing the size of the positive area would be greatest in the case of widely scattered thunderstorm activity, and least in an area of numerous, intense thunderstorms, such as along pronounced instability lines. There is no basic observational material on which to base a definite evaluation of the effect, however, and no operationally feasible procedure for including it at present.

Data from proximity soundings were used in the development of the SELS method so that, operationally, it is advisable to prepare a prognostic sounding representative of the centroid of a threat area in cases where no real sounding is sufficiently representative. As already stated, it is also necessary to make a prediction of thunderstorm occurrence within the threat area since the parameters utilized in hail size forecasting are not necessarily related to the probability of thunderstorm occurrence, beyond the indicated availability of the potential energy of hydrostatic instability.

The hail size is estimated from a measure of the positive area on a thermodynamic diagram (WB form 1147). This positive area usually approximates a triangle and the measurements used here are H (height of the triangle) and $\Delta T$ (base of the triangle in degrees C). Available data point to a temperature of -10°C as the mean level of hail formation. Thus, the height of the triangular positive area, H, is measured from the level of free convection (LFC) to the -10°C level on the predicted parcel lapse rate. The base of the triangle, $\Delta T$, is the temperature difference between the parcel and its environment. That is, the algebraic difference, -10°C minus the temperature of the actual environment or forecast sounding (forecast environment), at this same level. For this value to be significant, the parcel must be warmer than the environment. In those cases where the positive area is not approximately a triangle due to irregular observed or predicted temperature lapse rate, a smooth line is drawn to represent the mean lapse rate (equi-areal bisection). Actually, in working with prognostic soundings, such cases would not usually appear.

An overlay, shown in figure 38, is used to evaluate the positive area. This overlay is placed with the zero point (upper right) coincident with the -10°C point on the parcel temperature curve and the coordinates parallel to the appropriate coordinates on the WB 1147 A, B, or C base. An initial hail size estimate is read at the intersection of the isotherm corresponding to the environment temperature at the -10°C parcel temperature level and the isobar corresponding to the level of free convection. If the level of the -10°C parcel temperature is different from 400 mb (and it usually is) a correction for air density effect is made from the chart shown in figure 39, which is based upon calculations using a -10°C parcel temperature at 400, 500, and 600 mb. The intersection point of the hail size initially obtained from the overlay (entered along the sloping solid lines) and the actual pressure of the -10°C parcel temperature level (entered along the ordinate) is extended vertically to the top of this graph to obtain the corrected hail size. It will be noted that the higher the actual pressure of that level, the greater the hail size.

An example (graphical) of the use of this technique on a hail proximity sounding is shown in figure 40.

## EXTREME TURBULENCE

Extreme turbulence may be visualized as a field of intense vertical drafts and eddies of such dimensions, orientations, and strength that aircraft traversing the field are subjected to stresses that may do structural damage. The effective gust (i.e., the velocity of a "sharp-edged gust" that would have produced a normal acceleration equal to that experienced by the airplane) [12] has been utilized as a defining parameter since it may be used to estimate the effect that may be produced on any given aircraft by turbulence.

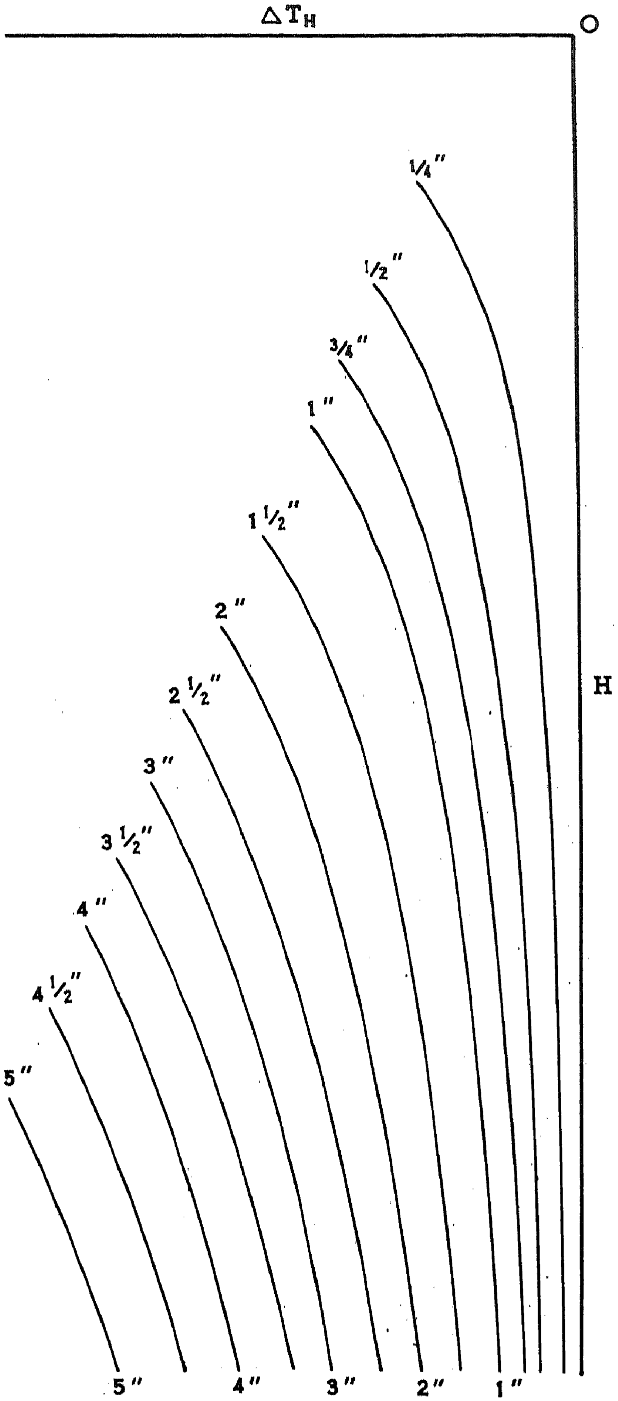

Figure 38. Overlay for pseudoadiabatic chart for estimating hail size. Zero point is placed at -10°C point on parcel temperature curve with the abscissa ($\Delta T_H$) along the pressure level of the -10°C intersection point and the ordinate (H) along the -10°C isotherm. Hail size estimates made from this figure must be corrected in figure 39 if the level of the -10°C parcel temperature is different from 400 mb. This figure is scaled so that it may be traced and used directly as an overlay on WB Forms 1147 A, B, or C.

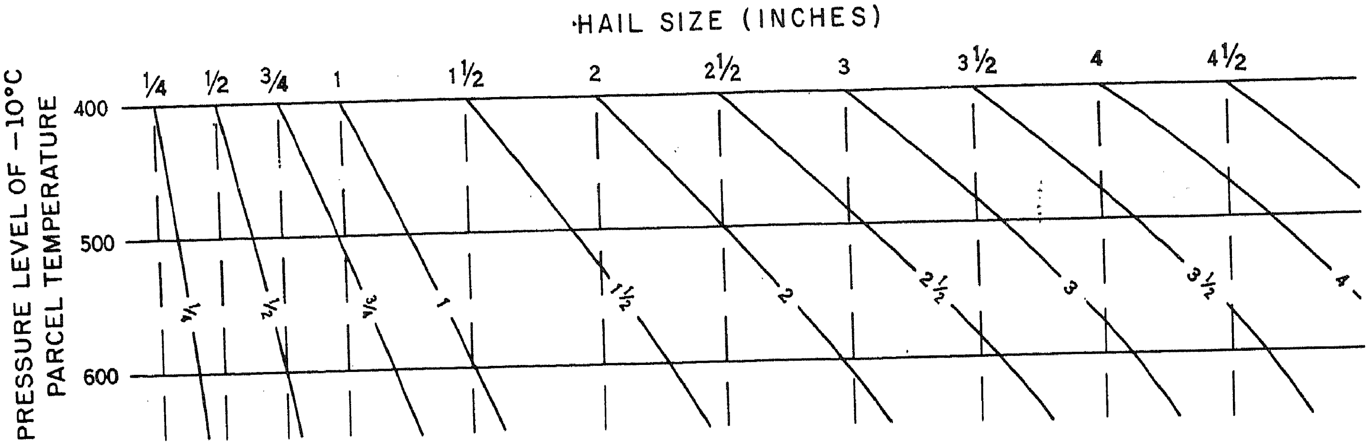

Figure 39. Correction graph for cases where parcel temperature of -10°C occurs at a level other than 400 mb. This graph is entered along the sloping line corresponding to the hail size obtained from figure 38. The corrected size is read vertically above the intersection of this sloping line and the pressure of the -10°C parcel temperature. Note that the higher the pressure the greater the hail size.

A technique developed by Bates [4] provides the basis of turbulence forecasts issued by the SELS Center. It was derived under the following assumptions: a) a definition of turbulence in terms of effective gusts; b) a correlation between updrafts and effective gusts of 0.42 as given in The Thunderstorm [12]; and c) a correlation between positive areas, as evaluated on pseudoadiabatic diagrams, and updrafts (In the hail size forecasting technique, the theoretically derived updrafts seem to be reasonably well related to positive areas).

The effective gust velocity which commercial aircraft are designed to withstand is 40 fps when flying at the rough air speed and 30 fps when flying at the design cruising speed. The turbulence predicted by SELS is that characterized by effective gust velocities in excess of 30 fps (updraft velocities greater than 80 fps), or those gusts exceeding the design specifications of most commercial aircraft in operation today.

The equation for the buoyant force upon an air parcel was used to plot the isopleth shown in figure 41, where the ordinate is in feet above the level of free convection and the abscissa is in degrees Celsius. This is used operationally as an overlay and permits a quick measure of the positive area which is, of course, related to updrafts. (It so happens that an updraft of 80 fps is that required to sustain a 3/4 inch hailstone where the -10°C parcel temperature is at the 400-mb level (see fig. 40).)

In using this overlay, the sounding (either actual or prognostic) is first analyzed as described under "Analysis of Soundings." The overlay is then placed so that the base corresponds to the

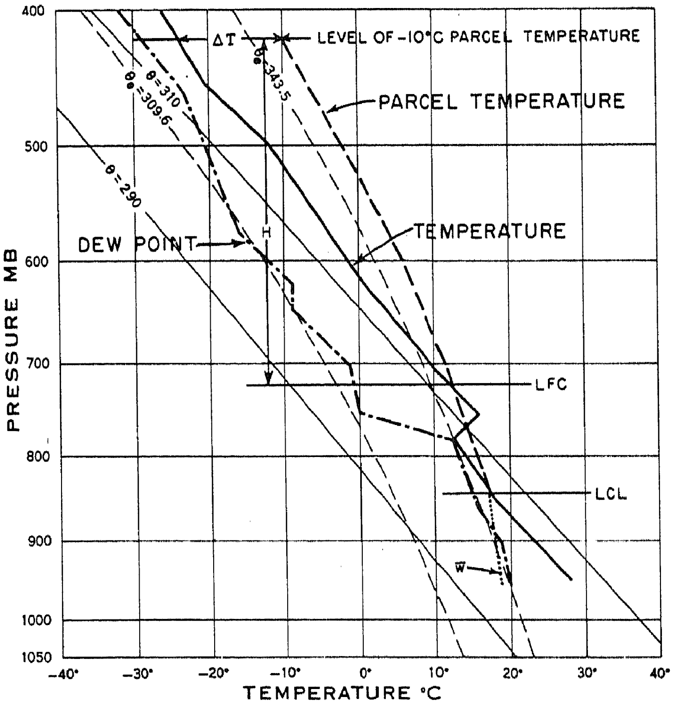

Figure 40. Upper air sounding at Altus, Okla., April 29, 1954, 1500 CST. Hail size was computed to be 2 1/4 inches. Hail up to 2 inches diameter fell 95 miles north of Altus, Okla. between 1530 CST and 1830 CST.

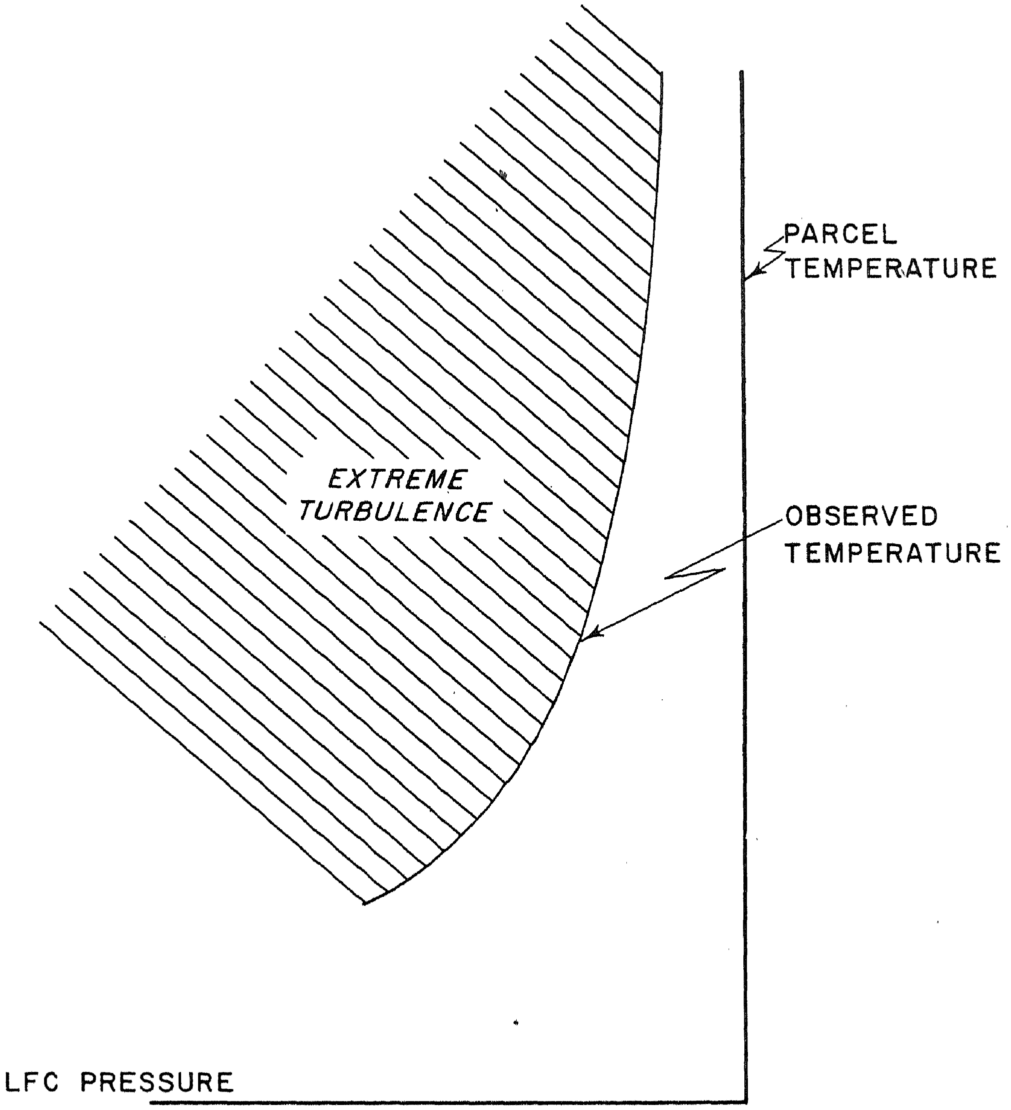

Figure 41. Overlay for pseudoadiabatic chart for determining extreme turbulence. This overlay is placed so that the base lies along the pressure line at the level of free convection. The overlay is then moved horizontally along the isobar so that the vertical axis intersects the parcel temperature curve at each level. Wherever the distance between the parcel temperature curve and the environment lapse rate is large enough to extend into the shaded zone, extreme turbulence is indicated. This figure is scaled so that it may be traced and used directly as an overlay on WB Forms 1147 A, B, or C.

pressure at the level of free convection. Then the overlay is moved horizontally along this isobar so that the vertical axis intersects the parcel temperature curve at each level. As the overlay is shifted the temperature differences between the environment (observed) and parcel on the pseudoadiabatic diagram are compared with the temperature differences necessary to define extreme turbulence as shown on the overlay. If the temperature difference on the pseudoadiabatic diagram, as analyzed, equals or exceeds the temperature difference indicated by the overlay, extreme turbulence is called for.

Usually no restriction on the height of extreme turbulence is made, but the more extreme turbulence is usually found in the higher portion of the sounding. In cases where very stable air is indicated at low levels, "extreme turbulence aloft" will be forecast.

An example of an analysis for turbulence is shown in figure 42.

## THUNDERSTORM SURFACE GUSTS

Efforts to employ the empirical method described by Fawbush and Miller [17] or to develop

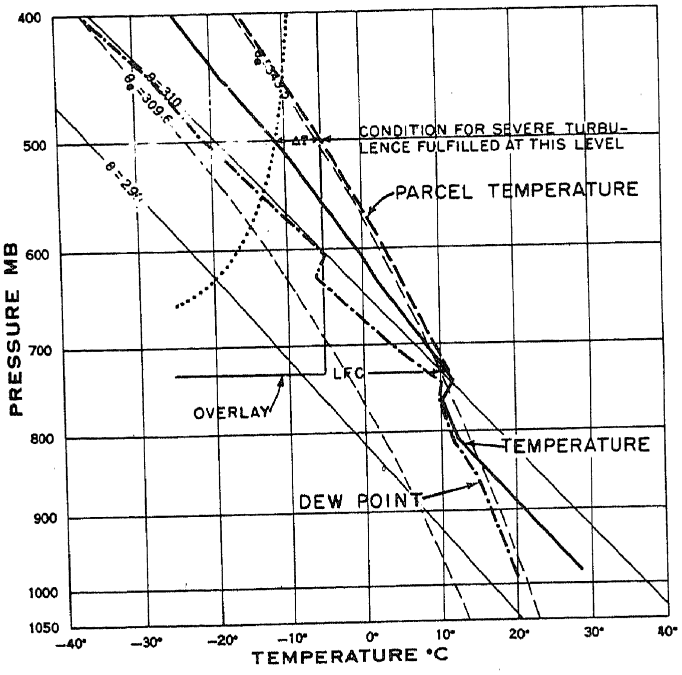

Figure 42. Upper air sounding at Fort Smith, Ark., Oct. 11, 1954, 1500 CST illustrating an analysis for severe turbulence. The overlay lines are drawn on this sounding and the condition for severe turbulence, as defined, is fulfilled at 18,000 feet. Severe turbulence was reported on the Oklahoma City-Little Rock route at 1603 CST between 18,000 and 28,000 feet msl.

a better method have met with so little success thus far that no attempt is made at this time to predict this particular feature of severe thunderstorms. While it is currently assumed that gust velocities are proportional to the intensity of the associated thunderstorm, it is well known that there are many exceptions. Research on this problem is continuing.

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